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THE CHANNELED SCABLANDS OF THE COLUMBIA PLATEAU

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DEFINITION OF "SCABLAND"

The terms "scabland" and "scabrock" are used in the Pacific Northwest to describe areas where denudation has removed or prevented the accumulation of a mantle of soil, and the underlying rock is exposed or covered largely with its own coarse, angular débris. The largest areas of scabland are on the Columbia Plateau in Wash-

ington, north of Snake River. These scablands have a history which is believed to be unique. The prevailing feature of their topography is indicated in the term here used: channeled scablands.[1] They are scored by thousands of channels eroded in the underlying rock. The plateau in Washington, north of Snake River, has a total area of about 12,750 square miles, of which at least 2,000 square miles is channeled scabland. The scabland is widely distributed over the region in linear tracts among maturely dissected hills which bear the loessial soil (wheat lands) of the plateau.

PHYSIOGRAPHIC RELATIONS OF THE CHANNELED SCABLANDS

The following features and relations of the scablands exist in all tracts. They must form the basis of any interpretation for the origin of channeled scabland. The map should be examined as this list is read.

1. Scabland tracts are developed invariably on or in the Columbia basalt formation.

2. Scabland tracts are invariably lower than the immediately adjacent soil-covered areas.

3. Scabland tracts are invariably elongate.

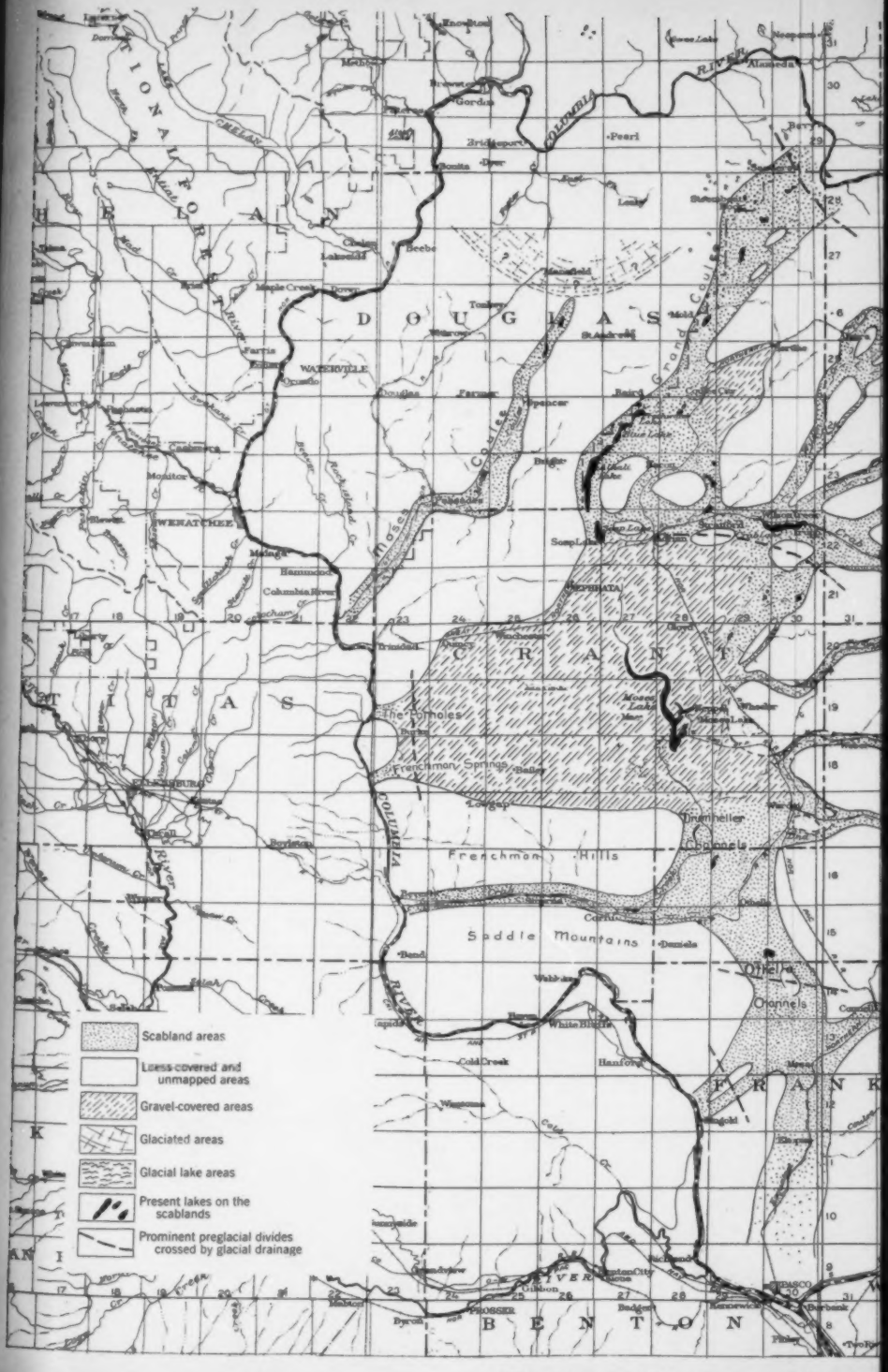
4. The elongation of scabland tracts is with the dip slope of the underlying basalt flows. There are eight known exceptions to this rule,[2] all minor affairs so far as length is concerned.

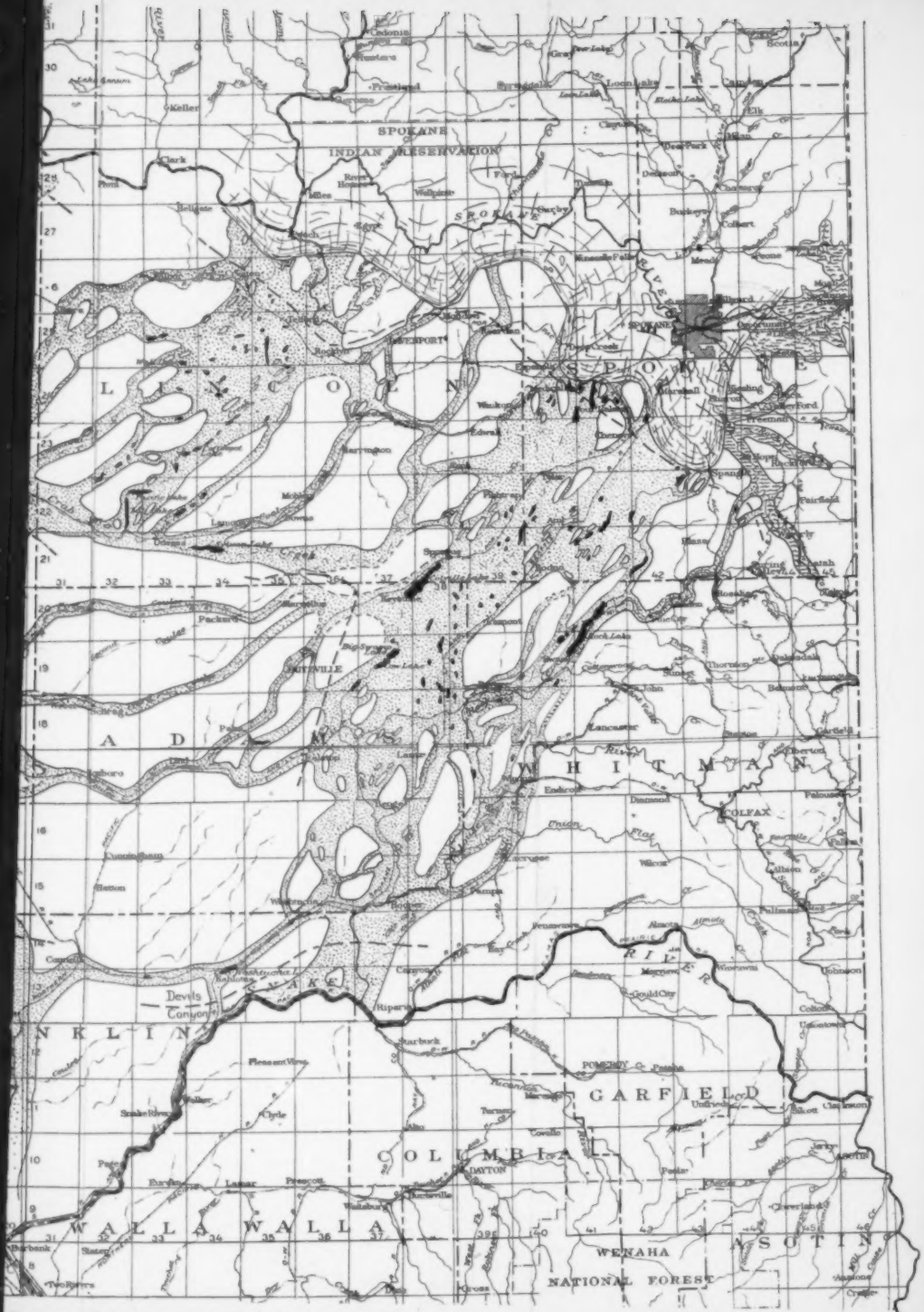
5. Scabland tracts, considered as units, invariably have continuous gradients.

6. Scabland tracts are invariably bounded by maturely eroded topography.

[1] An earlier paper on this subject was published by the writer in the *Bulletin of the Geological Society of America*, Vol. XXXIV (1923), pp. 573–608. The study on which it was based involved about a 1,000-mile traverse of the plateau. Since then, the writer has studied the plateau more thoroughly, having added more than 2,000 miles to the previous total traverse. Much more detailed information and several modifications of the earlier interpretations justify the appearance of a second paper on the subject. The accompanying map (Plate III) is based on a field examination of every scabland there indicated. In a few places the boundaries are inferred (dashed lines) but future work will hardly do more than make minor changes or additions.

[2] A part of Othello Channels, a part of Drumheller Channels, at Palisades and near Spencer in Moses Coulee, at Soap Lake and near Bacon in Grand Coulee, 6 miles south of Almira on Wilson Creek, and at Long Lake in Spring Coulee.


Scabland areas
Loess-covered and unmapped areas
Gravel-covered areas
Glaciated areas
Glacial lake areas
Present lakes on the scablands
Prominent preglacial divides crossed by glacial drainage
Columbia River
Grand Coulee
Moses Coulee
Frenchman Hills
Saddle Mountains
The Channels
Frenchman Springs
Drumheller Channels
Othello Channels
Cold Creek
Hanford
Wahluke
Pasco
Kennewick
Richland
Benton
Franklin
Douglas
Grant
Chelan
Okanogan
Ferry
Stevens
Lincoln
Adams
Yakima
Kittitas
Wenatchee
Ephrata
Soap Lake
Moses Lake
Quincy
Ellensburg
Othello
Pasco



COLUMBIA PLATEAU AND THEIR ASSOCIATED FEATURES

7. Scabland tracts are developed in pre-existing drainage lines of the mature topography.[1]

8. Scabland tracts are connected with each other.[2]

9. (*a*) The areas surrounded by scabland invariably have the dendritic drainage pattern, mature topography and loessial soil of the plateau. (*b*) They are almost invariably elongate with the scabland tracts. (*c*) They commonly have steep marginal slopes descending from 50 to 200 feet to the scabland. These slopes are almost invariably in loess. Slopes of 30° to 33° are not uncommon. They are much younger topographically than the slopes of the valleys among these mature hills.

10. Scabland tracts with steep gradient are narrow, while those with gentle gradient are wide.

11. The pattern of scabland tracts, where hills of the older topography are isolated in them, is anastomosing or "braided."

12. Scabland tracts invariably contain "channels." These are gorges or canyons or elongated basins eroded in the basalt. The channels are invariably elongate in parallelism with the tract as a whole and, in most cases, the channel pattern is anastomosing or braided.

13. (*a*) Scabland tracts invariably bear discontinuous deposits of basaltic stream gravel. (*b*) These deposits invariably contain a small proportion of pebbles and cobbles of rock foreign to the plateau. (*c*) These deposits invariably rest on an eroded, scabland surface of the basalt. (*d*) They commonly lie on the down-gradient side of eminences in the scabland.

14. (*a*) Scabland tracts invariably bear scattered bowlders of foreign rock. (*b*) The proportion of foreign débris, either the fragments in the gravel or the scattered bowlders, is invariably smaller with increasing distance down-gradient.

15. Scabland tracts are invariably without a mantle of residual soil.

16. Scabland tracts are traceable up-gradient to a narrow basalt plain bordering the south side of Spokane River in the northern

[1] There are many exceptions to this rule, occurring where scabland tracts cross pre-existing divides, but the total length of such is only a small fraction of the aggregate length of all scabland tracts.

[2] Moses Coulee is the only exception.

part of the plateau.[1] This basalt plain bears many glaciated erratic bowlders and some patches of till, but no channeled scabland, no mature topography, and no loessial soil.

17. Only where the minor valleys of the mature topography adjacent to this basalt plain open northward on to the plain are any glacial erratics found in them.[2]

18. There are but ten scabland openings to this basalt plain to the north.

19. Scabland tracts are invariably traceable down-gradient to Snake River on the south or to Columbia River on the west. There are nine places where scabland tracts enter these two streams Only three of them were drainage ways before the scablands were formed.

20. There is no channeled scabland on the plateau in western Idaho or south of Snake River or west of Columbia River.

21. Nowhere in the scablands or the maturely dissected country, during ten weeks of field study, has a till been found, or any deposit of doubtful genesis which could be interpreted more satisfactorily as till than as non-glacial in origin.[3]

GENERALIZED STATEMENT OF THE ORIGIN OF THE CHANNELED SCABLANDS

This unique combination of topographic features of the Columbia Plateau in Washington has only one interpretation consistent with all the foregoing items. The channeled scablands are the erosive record of large, high-gradient, glacier-born streams. The basalt plain records the southern limit reached by the ice sheet from which these streams took origin. Before this glaciation occurred, the entire plateau of Washington was covered with a loessial soil, varying in depth from a few feet to 200 feet. This and the underlying basalt had been eroded to maturity and a network of drainage lines covered it. The major water courses of this mature topography were consequent on the warped surface of the

[1] Grand Coulee and Moses Coulee are exceptions. Grand Coulee opens up-gradient into Columbia River Valley and the upper end of Moses Coulee is obliterated by the drift of a later glaciation.

[2] Exceptions due to a later episode in the history of the region are noted later.

[3] One exception, noted later.

plateau which descends in a general way from the northeast to the southwest.

The ice sheet approached and invaded the plateau from its northern high margin. It barely crossed to the headwaters of the consequent drainage. In places, it did not cross, but by blocking all other escapeways, its waters were forced to cross. By about a dozen different routes, at different altitudes and distributed along more than 150 miles of the ice front, water entered the mature drainage system. The capacity of the pre-existing valleys was wholly inadequate for the volume of most of these streams. Furthermore, gradients were high and the glacial waters eroded enormously, sweeping away the overlying loessial material, crossing low divides and isolating many groups of the maturely eroded hills to produce the anastomosing pattern of the scablands, biting deeply into the basalt to make the canyons and rock basins, and spilling into the Snake and Columbia in three times as many places as the pre-existing drainage had used.

This procedure of glacial streams was unique, so far as the writer is aware. It was unorthodox, at any rate, for no valley trains and but two outwash plains[1] were built on the plateau south of the basalt plain. The stream gravel of the scablands is almost wholly in separate bars.

The conception above outlined is amply sustained by every feature and relationship of the scablands. All other hypotheses meet fatal objections. Yet the reader of the following more detailed descriptions, if now accepting the writer's interpretation, is likely to pause repeatedly and question that interpretation. The magnitude of the erosive changes wrought by these glacial streams is nothing short of amazing. The writer confesses that during ten weeks' study of the region, each newly examined scabland tract reawakened a feeling of amazement that such huge streams could take origin from such small marginal tracts of an ice sheet, or that such an enormous amount of erosion, despite high gradients, could have resulted in the very brief time these streams existed. Not River Warren, nor the Chicago outlet, nor the Mohawk channel,

[1] The Hartline gravel flat and the Quincy basin fill, both in structural depressions in the plateau.

nor even Niagara falls and gorge itself approach the proportions of some of these scabland tracts and their canyons. From one of these canyons alone[1] 10 cubic miles of basalt was eroded by its glacial stream.

THE BASALT PLAIN, NORTH OF THE SCABLANDS AND MATURE TOPOGRAPHY

This physiographic feature extends westward from Spangle for 50 miles along the south side of Spokane River and varies from 3 to 12 miles in width. It is determined by the upper surface of the Columbia basalt formation. It is interrupted in places by short valleys tributary to Spokane River, and the different portions are known as prairies.[2] This plain is bounded on the south by channeled scabland and maturely eroded loessial hills. The differences between it and adjacent broader scablands are not marked, but the loessial hills are in striking contrast with it. These hills which, elsewhere on the plateau, come right to the edge of Snake and Columbia valleys,[3] nowhere overlook the Spokane Valley. They terminate abruptly on the southern margin of this plain. On the plain there is no mature topography and no channeled scabland. There is no area on the plateau like it. The nearest approach to it is the northern portion of Douglas County, back of the Wisconsin terminal moraine. This narrow plain must be the result of conditions which prevailed no farther south.

These conditions can be summed up in one word—glaciation. Deposits of till and many striated erratics have been found in every township examined on it. The till is patchy in distribution. No moraine margins the southern edge of the plain and no good moraine ridges occur on it, though here and there is morainic topography.

The genesis of the plain thus established, the questions of its character before glaciation and the method of its development arise. These are answered clearly when the adjoining mature loessial topography is studied. The larger valleys of this topog-

[1] Upper Grand Coulee.

[2] Paradise Prairie, Sunset Prairie, Indian Prairie, Four Mound Prairie, etc. On the north side of Spokane River are Pleasant Prairie and Five Mile Prairie, also parts of this plain.

[3] With the exception of northern Douglas County.

raphy have been eroded to varying depths into the basalt. The bottoms of such as lead out across the basalt plain to the north are lower among the hills than the general surface of the plain. The profile of the plain, extended back among the hills, cuts their slopes somewhere between hill summits and valley bottoms. And this transection is at the contact of loess on basalt. The ice sheet which covered the plain, therefore, simply removed the upper, weaker formation, and only to a minor extent altered the surface of the basalt formation.

THE MATURE TOPOGRAPHY

This is the dominant type of topography of the Columbia Plateau. The major drainage lines are structurally controlled, but the minor ones constitute a dendritic network. The pattern is eroded largely in a weak sedimentary deposit, chiefly loessial, which overlies the basalt. Maturity is expressed in the complete development of the drainage system, in the reduction of the original surface to valley slopes[1] and in the concavity of the lower part of many of these slopes. This maturity has been developed with reference to the underlying basalt as a base level, for progress of the cycle of erosion in the loess has been very much more rapid. Neglecting the loessial cover, the basalt plateau is in early youth, and will still be when the loess has been entirely removed. Nevertheless, the absence of deep trenches in basalt in the interior of the plateau, similar to Spokane, Columbia, and Snake River valleys about its margin, and the cutting through of the loess has allowed the development of shallow mature valleys in the upper part of the basalt.

The loessial deposit varies in thickness, in some places being only a soil, and in others being 200 feet or more in thickness. It is not all loess. There are places where it is chiefly a residual soil from the basalt, and others where it is a waterlaid sediment.[2] But many widely distributed sections show a succession of loessial deposits,

[1] A few broad, undissected divides are still left. Michigan Prairie, south of Lind, is a good example.

[2] Probably the Ellensburg formation. The Pleistocene Ringold formation, in Franklin County, is younger than the mature topography.

with abundant root and rootlet casts throughout and with reddened upper surfaces of each deposit, aggregating 50 feet or more.[1]

The mature topography is older than the scablands and the basalt plain. Literally hundreds of isolated groups of maturely eroded hills of loess stand in the scablands. Their gentle interior slopes are identical with those far from the scabland tracts. But their marginal slopes, descending to the scablands, commonly are very steep, over large areas amounting to 30° and even 35° (Figs. 1, 2, and 3). These steep slopes are seldom even gullied, except

FIG. 1.—One of a group of loessial hills in the scablands a few miles southwest of Rock Lake. One of the steepened slopes and its alignment are shown. Photo by O. C. Clifford.

where a drainage line leads out from the hill group to the scabland. Where the minor valleys transected by the steep slopes lead backward into the interior of a hill group they are simply hanging valleys.

There are few places where basalt occurs in these steepened slopes. Where present, it is always restricted to the lower part and shows itself in conspicuous ledges.

A very striking and significant feature of the steepened slopes is their convergence at the northern ends of the groups to form great

[1] Near Harrington and near Kahlotus are two excellent cuts which show this very well.

prows, pointing up the scabland's gradient (Figs. 2 and 3). The nose of a prow may extend as a sharp ridge from the scabland to the very summit of the hill. It is impossible to study these prow-

FIG. 2.—An isolated loessial hill on the scabland south of Hooper. The prow of the hill is pointed at the observer. The hill is 180 feet high, more than half a mile long, has a very narrow crest, and sides which slope 35°. It is entirely surrounded by scrubbed basalt. Half a mile to the west is a canyon in the scabland 75 feet deep, with an abandoned waterfall at the head.

FIG. 3.—The same hill as shown in Figure 2. The prow is at the right. The corresponding steepened slope at the tail of the hill shows at the extreme left. Most of the apparent left slope of the crest is a matter of perspective. Photo by O. C. Clifford.

pointed loessial hills, surrounded by the scarred and channeled basalt scablands, without seeing in them the result of a powerful eroding agent which attacked them about their bases and most effectively from the scabland's up-gradient direction.

DETAILS OF A SCABLAND SURFACE

All scablands are channeled to a greater or lesser extent. These channels are eroded in basalt to depths varying from a few feet to

FIG. 4.—Devils Canyon at mid-length, looking south. A double fall existed here when the canyon was eroded. The island and the eastern part still remain.

hundreds of feet. Commonly there are many shallow channels on each tract. Most tracts also have a few deeper channels, of the proportions of canyons[1] (Figs. 4 and 5). All channels in a tract are intricately interlaced, resulting in a multitude of butte-like hills, knobs, and ridges among them. Few channels have

[1] Upper Grand Coulee (1,000 feet deep), Lower Moses Coulee (900 feet deep), Devils Canyon, Franklin County (500 feet deep), and Palouse Canyon (500 feet deep) are the most noteworthy examples.

accordant grades where they unite or diverge, the bottoms of the shallower ones hanging above the floors of the deeper ones. Many canyoned channels have abandoned cataracts and cascades in them or at their heads.[1] Most canyoned channels have elongated rock basins (see Fig. 6). Even in the shallow channels, basins or pockets in the rock are common. Some of these rock basins clearly were produced by recession of a cataract whose scarp still

FIG. 5.—Devils Canyon near mid-length, looking north. Note the scrubbed basalt ledges above the canyon rim, and the profile of the loessial bluffs, still higher and farther back.

exists.[2] Others were produced by plucking of the columnar basalt in the canyon floors where the gradient was high.[3]

These features of the channeled scablands on the Columbia basalt plateau do not closely resemble any other type of topography.

[1] Dry Falls (400 feet high) in Grand Coulee, The Potholes (350 feet high) south of Trinidad, Frenchman Springs (400 feet high) south of The Potholes, and The Three Devils (600 feet total descent) in Moses Coulee are especially noteworthy.

[2] Deep Lake, below one of the Grand Coulee abandoned falls, has many associated huge potholes, drilled into the basalt at the foot of the falls as they retreated. Each of the two cataracts of "The Potholes," south of Trinidad, has a single elongated rock basin at the foot (Fig. 7).

[3] Rock Lake in Whitman County, Goose Lake in Grant County, Washtucna Lake and Eagle Lake in Franklin County, Pacific Lake and Tule Lake in Lincoln County, Goose Lake in Grant County, Medical Lake, Silver Lake, and Farrington (Fish) Lake in Spokane County, and Big Swamp Lake and Cow Lake in Adams County are examples of hundreds of such basins.

The narrowness and elongation of channels and the continuous gradients of tracts as a whole suggest river valleys but these features are all inherited from pre-existing valleys. Furthermore, some tracts are nearly as broad as they are long. The pattern of channels on a tract, like the pattern of some tracts and their isolated loessial hills, is much like that of great braided streams (Fig. 8).

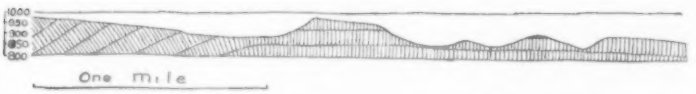


FIG. 6.—Longitudinal profile of the deepest of the Drumheller Channels across the nose of the anticline. Four rock basins are indicated, the largest of which is 75 feet deep.

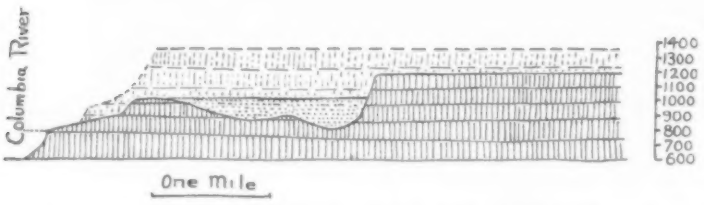


FIG. 7.—"The Potholes," longitudinal profile of the northern half, showing (1) cliff along northern side of the canyon, (2) amount of recession of the falls, (3) elongated rock basin below the falls, (4) great gravel bar along edge of the rock basin, and (5) approximate level of Columbia River when the cataract was formed.

But the scablands are erosional in origin, while the braided stream pattern is depositional.

The evidence for the origin of channeled scabland by stream erosion is overwhelming. The evidence of contemporaneity of action of all channels of a given tract, at least in its early history, is equally convincing. The only sequence indicated is that of development of the greater channels later in the epoch and consequent draining off of the shallower channels.[1] The scablands of the plateau in Washington are the beds of huge river courses in which

[1] Especially well shown in Lower Grand Coulee, in Moses Coulee between Palisades and Spencer, in The Potholes and Frenchman Springs south of Trinidad, and in Palouse Canyon south of Hooper, in all of which cataract recession in main channels cut off smaller channels alongside in the same scabland tract.

the streams once spread completely from side to side and only later became concentrated in the deeper canyons.

FIG. 8.—A part of Rock Creek (Lincoln County) and its associated scabland. The creek in the main channel is margined by vegetation. The scabland to the left is barren and the irregularity in lighting is due almost entirely to the cliffs which border the shallow anastomosing channels. (Aeroplane Photo by F. H. Frost.)

ALTITUDES AND GRADIENTS OF THE SCABLAND TRACTS

The heads of scabland tracts which are open to the basalt plain range in altitude between 2,350 and 2,500 feet above tide. There are six or eight of these, the number depending on just what is considered to constitute an individual scabland tract. At least the following should be recognized as unit tracts. The order in the list is from east to west.

1. Pine Creek channel. Altitude of head, 2,450 feet A.T.
   Gradient approximately 25 feet per mile
2. Cheney-Hooper tract
   Four heads:
   a) Marshall-Spangle, altitude, 2350±
   b) Four Lakes, altitude, ?
   c) Medical Lake, altitude, 2,425
   d) Deep Creek, altitude, 2,350±
   Total width of the group is 22 miles

Gradients:
Spangle to Hooper, 21.9 feet per mile
Cheney to Hooper, 22.6 feet per mile
Medical Lake to Sprague, 24 feet per mile
Medical Lake to Hooper, 22.5 feet per mile

3. Reardan channel. Altitude of head, 2,500+
Gradient, Reardan to Odessa, 20 feet per mile
4. Davenport-Harrington channel. Altitude of head, 2,450±
Gradient:
Davenport to Harrington, 27 feet per mile
Harrington to Odessa, 26 feet per mile
5. Telford tract
a) Eastern head. Altitude, 2,500±
Gradient:
Rocklyn to Odessa, 30 feet per mile
Rocklyn to Krupp (Marlin), 26 feet per mile
b) Western head. Altitude, 2,500±
Gradient:
Near Creston to Krupp (Marlin), 32 feet per mile
Near Creston to Wilson Creek, 30 feet per mile
Total width of heads is 17 miles
c) Wilbur branch
Gradient:
Creston to Wilson Creek, 32 feet per mile
Wilbur to Wilson Creek, 25 feet per mile
Almira to Wilson Creek, 26 feet per mile

The only scabland tracts which do not open on the basalt plain are Grand Coulee and Moses Coulee. For the head of Moses Coulee there are no altitude measurements. Grand Coulee has had a peculiar history, not yet fully deciphered, but the significant altitude at its head for present purposes is not the coulee floor (1,530 feet A.T.) but the scabland margining the brink of the canyon, about 2,500 feet A.T. The canyon has been cut subsequent to the first spilling over of glacial waters. The floor near Coulee City is 1,510 feet A.T. Most of this descent occurred within a few miles of Coulee City, the original slope being as steep as 20° in part and averaging perhaps 10° for 1,000 feet of descent. This is the chief reason for the great canyon across the divide. No other scabland head has been notably canyoned. None other had a gradient to exceed about 30 feet to the mile. All the canyons of the channeled

scablands are located in places of exceptionally steep original gradients.

If the channeled scablands are the product of stream erosion, and if all tracts are of the same age, the fact that the scabland heads vary in altitude, though all but two are open to the same glaciated basalt plain, can have but one satisfactory explanation. Each glacial stream must have had a source of water which was unconnected with the others. This means that the ice sheet whose melting yielded these streams must have covered the basalt plain and in most places must have been in contact with the mature loessial hills which separate the scabland heads. It means, furthermore, that but a few miles of ice front supplied the water for streams so huge that they flooded over many preglacial divides of the mature topography even in the southern part of the plateau. It was this flooding across divides which produced the scabland plexus of the plateau.

About 40 miles of ice front in one case[1] yielded water sufficient to denude a non-elongated tract 250 square miles in area of a loessial cover about 100 feet in maximum thickness. This was done by lateral planation in the preglacial drainage lines of the tract. These lines were so shallowly intrenched in the basalt and the volume of the water was so great that, as the loessial hills were eroded away, the flood spread over the entire area, 13 miles wide. Gradients were low, however, and it did not develop canyoned channels. Steeper gradients farther from the edge of the ice, and greater capacity of the preglacial valleys, held the waters within the confines of these valleys, but six such[2] were necessary to contain the flood and they were all greatly eroded in the underlying basalt.

Another large scabland area[3] is 75 miles long and 15 miles in average width. Its total descent is 1,850 feet. Its altitude at the head is the lowest of all such tracts. It differs from the one above in its notable linear extent, in the possession of a large number of isolated groups of loessial hills of the older topography, in its greater

[1] The Telford scabland and its dependencies.

[2] Coal Creek, Duck Creek, Lake Creek, an unnamed creek, Connawai Creek and Wilson Creek.

[3] The Cheney-Hooper area, extending from Spangle and Cheney to Snake River, south of Hooper.

gradient and in the development of canyons.[1] Though a much greater volume of water passed through this tract, the gradients were steep enough to draw off the flood and prevent a complete spreading over the area. Much water came to it from the ice margin to the east in Idaho and no estimate can now be made of the length of ice front which contributed to it. There were five or six places of distributary discharge where this flood crossed preglacial divides and one of them[2] eventually obtained most of the discharge. Along this distributary route, the glacial flood swept away the loessial hills for a width of 10 to 15 miles and eroded 500 feet into the basalt.

The glacial drainage route which possessed the highest crossing of the preglacial surface of the plateau is Grand Coulee. It also found the steepest gradient[3] and its volume was sufficient with this gradient to cause the deepest erosion in the basalt. Upper Grand Coulee, across the preglacial divide, is 1,000 feet deep. But its floor, after the epoch had closed, was lower than that of any other glacial drainage route at the head. In its early history it drew water from about 40 miles of ice front, but never spread widely. The gradient, determined here by exceptional warping of the basalt, prevented that. There was no noteworthy preglacial stream along its course. Grand Coulee is, therefore, the simplest but grandest case of canyon-cutting by glacial streams on the plateau.

DEPTH OF GLACIAL STREAM EROSION IN THE SCABLANDS

Criteria.—The courses of the larger valleys in the mature drainage system were determined by the warped surface of the basalt. The dominant feature of this warping is the southwestward dip from Spokane River and Columbia River on the north to Snake River on the south and Columbia River on the west. Many minor folds are superposed on this dip slope, so recent geologically that anticlines determine divides and synclines contain stream valleys.[4] Commonly, only the major valleys are intrenched in the

[1] Cow Creek, Rock Lake, and Creek and lower Palouse River now occupy the most striking of these canyons.

[2] Palouse River Canyon below Hooper.

[3] In one place, 1,000 feet in about a mile.

[4] Examples are Moses Coulee east of Palisades, Wilson Creek above Almira, Crab Creek below Corfu, Washtucna Coulee, Lind Coulee below Lind, Snake River near Lewiston and Clarkston, Union Flat Creek, Rebel Flat Creek, and Palouse River above Winona.

basalt. The minor valleys, in general, are not eroded through the loessial mantle.

Except near the bounding canyons of Spokane, Columbia, and Snake Rivers, or in exceptionally upwarped parts of the plateau, the preglacial valleys in basalt which were unviolated by the glacial flood, have the same mature slopes as those in the loess. Where glacial streams found routes eroded but slightly in basalt, they commonly spread widely at first[1] and only later eroded canyons, if they did so at all. Such wide scablands commonly are bounded by steep bluffs of loess. Many of the mature valleys in basalt were sufficiently capacious to contain the glacial waters which entered them. In such cases[2] the scabland of the route lies on the sides and bottoms only, and unsteepened slopes of the bounding older topography may come to the edge of the scabland. Such a tract is narrow and instead of having a multitude of lateral shallow channels anastomosing with the main canyon, it consists of exceedingly roughened ledges of basalt outcropping on the slopes of the main valley. Shoulders in the curves of these valleys were treated with especial vigor, in some cases being wholly cut away to leave prominent cliffs on the valley walls.

By smoothing out these ledges, something of the cross-section of the preglacial valley may be obtained. Remnants of the old floors are recorded in isolated buttes of basalt on the present floor and in the lowest prominent rock terraces, below which is the canyoned channel eroded in mid-current by the huge glacial stream.

Instances.—In Cow Creek, southwest of Ralston, the remarkable number of knobs and buttes in the lower part of the valley indicates clearly that the preglacial floor must have been at least 75 feet higher than the present. In Crab Creek Valley, between Krupp (Marlin) and Stratford, there are prominent buttes, isolated or partially isolated on the floor of the canyon. The tops of these are remnants of the preglacial floor and their height (100 feet in places) is a minimal measure of the depth of canyon-cutting by the glacial waters.

[1] As in the Telford and the Cheney-Hooper areas.

[2] Washtucna Coulee, Esquatzel Coulee, Lower Moses Coulee, Lind Coulee, Pine Creek above Rock Lake, Rock Creek (Lincoln County) and Coal Creek are examples.

In Washtucna Coulee, numerous prominent rock terraces from 150 to 200 feet above the present floor are probably remnants of the earlier valley bottom. In Esquatzel Coulee, into which Washtucna opens, these terraces are 200 feet and more above the bottom.

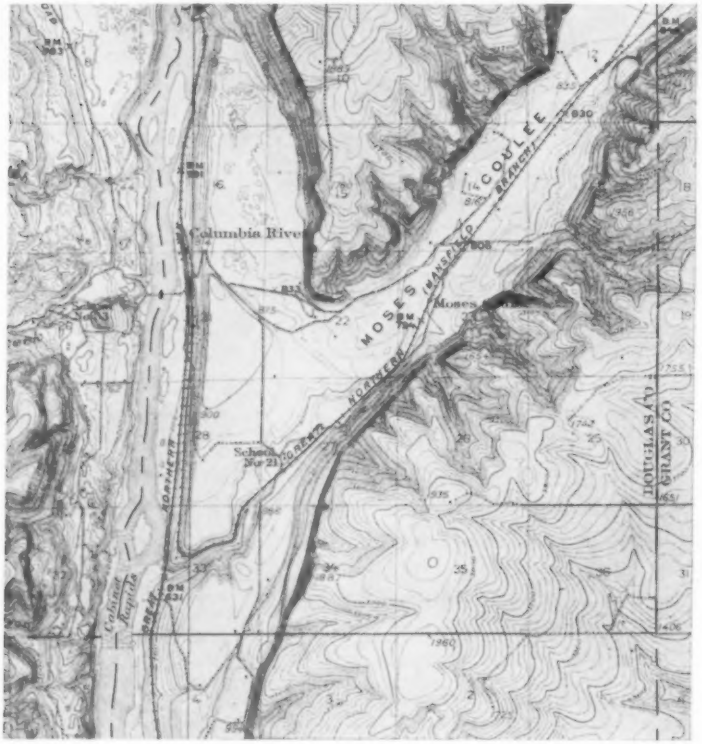


FIG. 9.—The lower part of Moses Coulee. Part of Malaga, Washington, topographic map. Note truncated spurs, hanging ravines, alluvial fans, and the great terrace in both Moses Coulee and Columbia Valley.

In Lower Moses Coulee (Fig. 9), the mouths of preglacial tributary ravines hang approximately 400 feet above the rock floor. Some of this discordance of grade is due to widening of the preglacial coulee, which here was a canyon, by the glacial stream but probably most of it is due to deepening during the glacial epoch.

In addition to the deepening of valleys already existing, the glacial flood actually made, *de novo*, drainage ways of greater width and depth than any previously developed on the plateau. This happened where divides were crossed and unusually high gradients down the farther slopes were found. Five such places are especially noteworthy: Devils Canyon, Palouse Canyon below Hooper, Drumheller Channels, Othello Channels, and Grand Coulee.

Preglacial Palouse River joined Snake River at Pasco, its subparallelism with the larger stream for 150 miles being structurally determined. The glacial flood from the north entered it in mid-length at several places between Winona and Washtucna. The volume of this flood was more than the valley could carry away.

FIG. 10.—Devils Canyon, cross-section a mile and a half south of Kahlotus, showing (1) steepened loessial bluffs (33°), (2) narrow scabland above brink of basalt cliffs (3) canyon 450 feet deep and less than one-fourth mile wide, and (4) post-Spokane talus, three-fourths the height of the cliffs. Horizontal and vertical scale the same.

Two leaks across the divide to the Snake developed, one near Kahlotus, and one near Hooper, and in both very great gradients were encountered.

The Devils Canyon distributary, south of Kahlotus, cut 50 feet or so through the loess and then by recession of waterfalls over ledges of basalt in the north slope of Snake River Valley, it eroded a canyon 5 miles long, a quarter of a mile wide and 500 feet deep (Fig. 10). Every fall but the lowest retreated completely through the divide. The remaining ledge, like a dam separating the two canyons, is less than 100 feet above the floor of Washtucna Coulee and not half a mile wide.[1] An abandoned half of a double cascade stands in mid-length of Devils Canyon (Fig. 4).

[1] The Spokane, Portland and Seattle Railroad tunnels through this ledge. In 1916, Washtucna Lake was so high for a time that it overflowed through this tunnel into Devils Canyon.

A much larger volume of water spilled across the preglacial divide south of Hooper. Before the great canyoned channel now transecting this divide had been eroded, the stream was 10 to 15 miles wide. Several channels of canyon proportion were initiated, but one outran the others in its deepening and finally secured all the discharge. Many of the shallower canyons enter the southern part of the main one over abandoned waterfalls. These channels could have carried water only when the wide scabland of the divide had no deep canyon completely across it. Their falls could have developed only after a deep main canyon existed *below* their junction. It follows, therefore, that Palouse Canyon was cut by retreating waterfalls, though these were destroyed later in the epoch. One only survived, now notched considerably by the post-glacial work of the Palouse River. The falls today are 198 feet high. Palouse Canyon is another Devils Canyon in all save its greater width and the fact that the preglacial divide was cut entirely in two.

Drumheller Channels and Othello Channels are two remarkable cases where the glacial flood crossed anticlinal ranges.[1] In both, the anticline is asymmetrical and the waters flowed down the gentle slope. In both, the flood at first was wide but became concentrated later in the deepening canyons. The maximum depth of erosion in Drumheller Channels was 400 feet, about 100 feet of which was in a weak sedimentary formation (probably the Ellensburg), and 300 feet in basalt (Fig. 11). The width of the Drumheller denuded tract is about 10 miles. This particular scabland area has a more striking and more complicated development of channels and rock basins than any other on the plateau. It is the only area of this type now topographically mapped. Below the Channels, along the northern flank of Saddle Mountains, the ancient river eroded 300 feet into the broader scabland.

But Drumheller Channels is not wholly the product of the glacial flood whose history we have been following. It and the main canyon of Grand Coulee carried drainage from a later ice sheet, and it has carried Crab Creek since the later glacial epoch. The amount of deepening during each Pleistocene epoch is diffi-

[1] Drumheller Channels crosses the eastern nose of Frenchman Hills anticline, Othello Channels crosses the eastern nose of Saddle Mountains anticline.

cult to determine. Othello Channels carried less water than Drumheller Channels, consequently is a smaller tract. Furthermore, it received water during only the earlier epoch. The degree of development of its canyons and rock basins is comparable to that of the larger tract and makes it probable that most of Drumheller Channels were formed during the earlier epoch.

The features of Grand Coulee are of such magnitude and its history so complicated by local conditions that an entire paper might well be devoted to it. It affords the greatest example of canyon-cutting by glacial streams, not alone for the Columbia Plateau, but for the world. The field evidence indicates that no preglacial drainage route ever existed here. Scabland with shallow channels margins the upper part of the Coulee, though 1,000 feet higher than the adjacent coulee floor, and there are no tributaries in the mature topography such as are possessed by Lower Moses Coulee and Washtucna Coulee. A glacial river, 3 miles in minimum width, spilled southward here over the divide and down a steep monoclinal slope. Judging by present grades and altitudes of this structural slope, the stream descended nearly 1,000 feet on a grade of approximately 10°, a few miles north of Coulee City. Such a situation is unparalleled, even in this region of huge, suddenly initi-

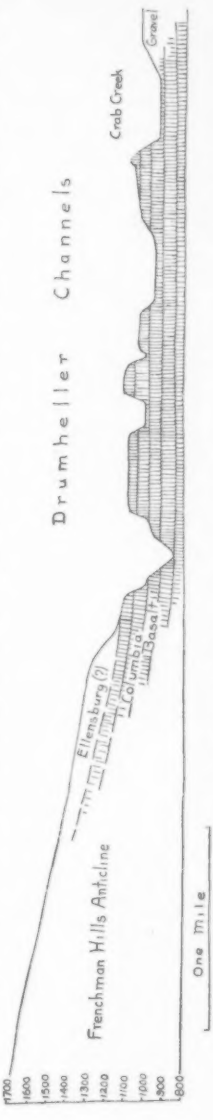


FIG. 11.—Cross-section of the head of Drumheller Channels, showing (1) structure at eastern nose of Frenchman Hills anticline, (2) the pre-scabland topography, and (3) the scabland channels. Probably no preglacial drainage crossed the anticline at this place. The lower 50 feet of the deepest channel is a rock basin.

ated, high-gradient rivers. Across this monocline between Columbia River and Coulee City, the canyon is 30 miles long and averages all of 2 miles in width and 800 to 900 feet in depth. In the making, at least 10 cubic miles of basalt were excavated and removed. Though a later flood of glacial waters used this route,[1] it did but little to deepen it.[2] By far the greater part of the erosion of Upper Grand Coulee was performed by the earlier glacial stream. It is very probable that this immense task was performed by the stoping of cascades and cataracts which retreated entirely through the monoclinal uplift to the deeper valley of Columbia River and thus left the great notch.[3]

It also seems probable that when the retreat of the ice sheet began, the plateau west of Grand Coulee was abandoned last. Earlier clearing of the Spokane and Columbia valleys to the east allowed all the drainage of the ice sheet to use the Grand Coulee route, which was then the lowest of all. Grand Coulee's greatest flood and probably its greatest erosion thus came after the other scabland routes had gone dry.

VOLUME OF THE GLACIAL STREAMS

If the channeled scablands of the Columbia Plateau are the erosive results of glacial waters, certain statements as to the volume of the streams can be made. Measurements are possible if remnants of the preglacial floor of the main valley exist in places where the valley brimmed over with the glacial flood to produce distributary courses. Should it appear that the amount of canyon-

[1] During the Wisconsin glaciation.

[2] Grand Falls, below Coulee City, consists of Dry Falls, Deep Lake Falls and a smaller unnamed falls a mile east of Deep Lake. The lip of the smaller falls is 125 feet higher than the floor of the channel leading to Dry Falls. Yet all were made by the same glacial stream and only Dry Falls and Deep Lake Falls were used and modified by the later discharge.

Furthermore, the Wisconsin ice did not cross Spokane River or Columbia River east of Grand Coulee and its waters were free to use the lowest of the ten earlier routes. Only Grand Coulee was so used, showing that it had been eroded by the earlier discharge to a depth not far short of that which it now has.

[3] This inference has no physiographic evidence in Grand Coulee to substantiate it, but is based on the known procedure of the glacial streams in similar situations, e.g., Lower Palouse Canyon, Devils Canyon, Frenchman Springs, and The Potholes.

cutting by these glacial streams has been overestimated, the depth of the stream to flood across the divide must be correspondingly increased. This view promptly runs into an absurdity, for the less the canyon-cutting is held to have been, the deeper and therefore more competent to erode the stream must have been.

One of the best cases for measurement is Washtucna Coulee at the head of Devils Canyon. Though there are but two small rock knobs out in the coulee floor, the summit of neither indicating the original valley bottom, there are good rock terraces to record it (Figs. 12 and 13). Near Kahlotus they lie 1,000 to 1,200 feet above tide. The glacial stream here, at the beginning of its history, overflowed through the loessial hills to Snake River, at an altitude of at least 1,350 feet. It was, therefore, from 150 to 350 feet deep. Its width averaged at least a mile.

And this was no ponded condition, for Washtucna Coulee opened widely into Esquatzel Coulee, an even more capacious valley, and Esquatzel in turn into Columbia and Snake valleys, and the glacial waters cut deeply into the bottom of both coulees. Figuring the preglacial floor as 1,000 feet A.T. at Kahlotus and as 675 feet A.T. at Eltopia, 34 miles farther down the valley, the great glacial stream had a gradient of about 10 feet to the mile.

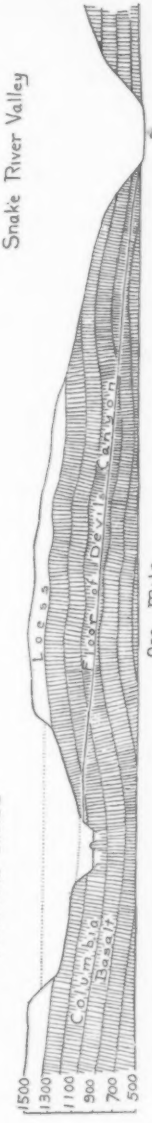


FIG. 12.—Cross-section of Washtucna Coulee and Snake River Valley, with longitudinal profile of Devils Canyon, showing (1) structure of the basalt, (2) its loessial cover, (3) rock terraces flanking Washtucna (Kahlotus) Lake, remnants of the preglacial valley floor, (4) the inner canyon eroded by the glacial stream, (5) steepened slopes of the loess, facing the coulee, and (6) approximate upper limit of the flood which spilled across to Snake River.

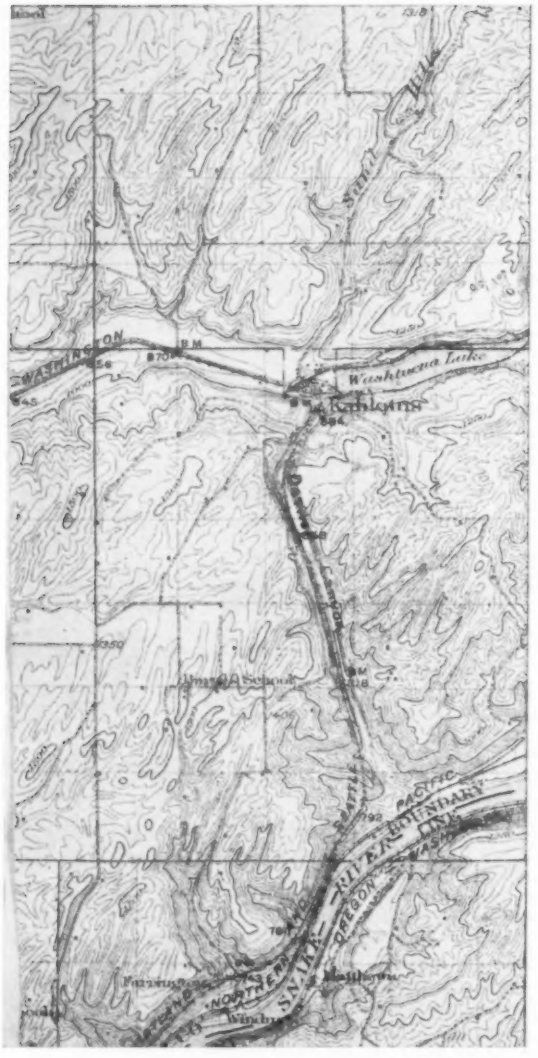


FIG. 13.—Devils Canyon. Part of Connell, Washington, topographic map.

Further evidence that glacial waters so filled Washtucna Coulee that the former valley became simply a channel is found in the upper limit of glacial stream gravels and of scablands. At the head of Devils Canyon, the highest scabland is 1,250 feet A.T., 250 feet above the brink of the cliffs of the canyoned channel. At Estes, a gravel bar deposited by the glacial stream lies 250 feet above the Coulee floor and about 125 feet above the rock terrace which marks the old valley floor. Near Sulphur, the highest scabland surface, at the base of the steep loessial bluffs, is between 1,100 and 1,150 feet A.T., 100 to 150 above the rock terrace. Northwest of Connell a terrace of sand and fine gravel lies at 1,000 feet A.T. It marks the upper margin of the scabland here and probably is a deposit of the glacial stream. It is 100 feet above the broad rock terrace. The canyoned channel here is cut 150 feet below the rock terrace.

Crab Creek Valley, below Odessa, received more water than it could carry away, at least before its central canyon had been eroded. It overflowed southwestward, by way of Black Rock Coulee and its associated scabland, to the Quincy Basin which Crab Creek itself entered farther north. Measurements here are only approximate but they indicate the order of magnitude of this glacial stream. Scabland and glacial stream gravel along the southern edge of Crab Creek Valley lie 300 feet or more above the present stream and extend a mile and a half back on the upland from the margin of the preglacial valley. This valley had been canyoned more than 100 feet by the glacial stream which, on the basis of these figures, was 200 feet deep at its inception.

The Telford scabland tract, 13 miles wide and 20 miles long, has been swept almost completely bare of the loessial deposit. The relief in a cross-section of the basalt surface now exposed, aside from the minor canyons, is about 50 feet. To have been so denuded, this tract must have had a sheet of running water of this depth completely over it.[1]

[1] That the ice sheet did not advance over the Telford denuded tract is shown by the presence of a few isolated loessial hills with characteristically steepened marginal slopes. One such group lies 7 or 8 miles north of Telford. It has a maturely eroded topography and a dark loessial soil without rock fragments of any kind. But it is cut by channels of glacial waters which eroded to the basalt. The fact that these waters went through the group, though the surface immediately north of them drops off into the deep canyons of Hawk Creek, a tributary of Spokane River, proves that glacial ice must have crowded up against the northern side of the group.

Further evidence of the depth of the glacial streams will be presented under the next subject.[1]

DEPOSITS MADE BY THE GLACIAL STREAMS

The record of Pleistocene glacial streams almost everywhere is one of aggradation. Glacialists commonly think of the subject only in such terms, textbooks discuss it only in that light; it is the orthodox conception. But in the Columbia Plateau exceptional factors controlled. The preglacial valleys in general were small and of relatively high gradient, the volume of the glacial streams was very great, the amount of detritus from the ice was very small, and the rock crossed was either loess or closely and vertically jointed basalt, both of which yielded rapidly to the torrents. The result of these conditions was great deepening of the valleys, and deposits made by the streams are of minor importance. Their character, however, adds to the weight of evidence already presented for the origin of channeled scabland by glacial stream erosion.

The deposits are almost wholly of gravel. Sand is a minor constituent and clayey material is lacking. The gravel and sand are almost wholly of basalt, though all deposits contain fragments of rock foreign to the plateau. The basaltic gravel is not well rounded though most of it is sorted and stratified and indubitably of stream origin. Foreset bedding is common, the direction of dip according with the slope of the scabland tract. In some places, the deposits are composed of very coarse material, with abundant, sub-rounded, basaltic bowlders 3 and 4 feet in diameter. These were originally bowlders of decomposition and were derived from flows with particularly large columns, underlying or in the immediate proximity of the deposit. Where a few erratic bowlders are associated[2] the deposit itself might be misinterpreted as a bowldery till.

The gravel deposits rest on irregularly eroded basalt, essentially a buried scabland surface. Nowhere do they lie on or beneath the loess. Neither the gravels nor the underlying basalt are

[1] If these enormous streams all came to the Columbia eventually, should not the great volume be recorded farther down the master stream? The writer has seen enough to convince him that it is so recorded, and hopes to publish on this subject in the future.

[2] As west of Lantz, for example.

decayed.[1] Cementation with calcium carbonate has begun but has not advanced far.

Most isolated loessial hills in a scabland tract have a deposit of gravel depending from their down-gradient end. Many knobs and buttes of basalt have similarly situated gravel deposits.

In many cases, the gravel deposits constitute discontinuous terraces on the margins of the scabland tracts, suggesting remnants of former valley fills.[2] But the evidence seems conclusive that all gravel deposits of the scablands are bars, built in favorable situations in the great streams which eroded the channels.

The rounded profiles and ground plans of many gravel deposits in the scablands are in accord with this interpretation. The unfilled canyons and rock basins, in intimate association with the discontinuous gravel deposits, indicate clearly that both are products of the same episode. The only alternative hypothesis is that channeled scabland was formed, then buried in gravel, then in large part re-excavated by streams little short of the magnitude of those which eroded the scablands. This has no other field evidence to support it and requires a much more complicated history. Furthermore, such deep canyons were cut when the scablands were made, and such noteworthy divide crossings were made that a reoccupation of all the scablands by glacial drainage from a second ice sheet would be impossible. And the hypothesis of dissection by the postglacial streams of the scablands is quite inadequate. Lakes and pools still stand in the rock basins on the channel floors, almost as they were left by the glacial flood.

Gravel deposits in the deeply canyoned scablands occur on the broad upper scabland surfaces, on the roughened slopes of the preglacial valleys and down in the canyons. The interpretation of these deposits as bars requires no change in general conditions, as does the alternative hypothesis; it simply requires that gravel be deposited locally as conditions might favor, all through the

[1] Local exceptions, as 1 mile southwest of Lamont, where ground water has been especially active.

[2] So the terraces in Pine Creek channel were interpreted in the earlier paper by the writer. That view is here abandoned for one much more consistent with all other features of the scablands.

epoch of erosion of the basalt. Deposits on the highest scablands[1] were made before the deepening canyons drew the waters into more restricted routes. Deposits in the deeper portions[2] were made during the latest stages of the episode. No gravel deposits were ever as thick as the rival hypothesis would require them to be.[3] That view demands that the canyons be eroded, then filled completely *with their own débris*, then re-excavated in large part.

However, there are two places on the plateau where the history of deposition by glacial streams has been somewhat different. One of these is Hartline gravel flat, the other is Quincy Basin. Both are structural depressions, not completely filled before the glacial floods arrived. The Hartline structural valley became filled with débris from the cutting of Upper Grand Coulee before Lower Grand Coulee had been eroded. The trenching of the lower coulee, and particularly the development and retreat of Grand Falls, incised the southern rim of the valley so that, by the close of the episode, the gravel fill, once the floor of the glacial stream, had been removed in its western part, and the remnant left 200 feet or so higher than the brink of Grand Falls.[4] The total fill in the Hartline structural valley is about 250 feet. It is composed of bowlders, cobbles, and gravel near Grand Coulee, and of sand 5 or 6 miles east, back from the main drainage line. An old channel from Grand Coulee crosses its northern and eastern part and leads into Deadmans Gulch, a distributary which spilled across the southern rim into Spring Coulee before Lower Grand Coulee had developed into the dominant notch that drained and dissected the flat. The terrace form of Hartline flat is the result of erosion of a once complete gravel fill. Its scarp is not constructional, as are those of bars. But the fill and the subsequent erosion occurred because of special local conditions, not because general conditions changed.

[1] As about Gloyd along the Black Rock distributary from Crab Creek Valley.

[2] As at the mouths of Duck Creek and Wilson Creek in Crab Creek Valley and at the junction of Crab and Coal creeks.

[3] For example, gravel deposits north of the town of Washtucna lie on the slopes of the coulee, 350 feet above the present floor. They antedate the deeper canyoned portions of the coulee.

[4] Estimated from the surviving eastern member of that complex waterfall, the only part which escaped modification by the Wisconsin glacial stream.

Quincy Basin, like the Hartline Valley, does not have a scabland floor. Both lay too low to be eroded. But both belong to the glacial drainage plexus. Quincy Basin probably contains more gravel than all the scablands of the plateau together. It was an enormous settling basin for the glacial rivers from Grand Coulee and Crab Creek. The flood of waters entering it was so great that at first three discharge ways were simultaneously in operation.[1] The southern and larger one obtained all the discharge later, and by deep notching of the basin's rim, caused the glacial waters which traversed the fill to incise their deposits. Two great channels and one smaller one were thus formed. The two large channels are each about 3 miles wide. Each was eroded about 100 feet deep during the later part of the episode. The one which contains Rocky Ford Creek also carried the later Wisconsin discharge and was further modified then.[2]

Erratic bowlders, some of them striated, are widely distributed at all altitudes on the basalt plain and the scablands. They also occur in valleys of the mature topography which open northward on to the basalt plain, and in some which open on to scabland tracts. The size, angularity, and striated surfaces indicate that these erratic bowlders were not rolled to their positions by running water. In the scablands, they must have been carried by berg ice on the great rivers. In their peculiar and limited distribution in the valleys in loess is evidence of small glacial lakes, in which the drift-bearing bergs floated.[3]

[1] Frenchman Springs, The Potholes, and Drumheller Channels.

[2] This interpretation is a modification of that published earlier by the writer. Further study of Grand Coulee, Quincy Basin, and Drumheller Channels has led to a magnification of the work of the earlier flood, and a minimizing of the results accomplished during the Wisconsin epoch. Grand Falls is now considered to be a pre-Wisconsin affair, none of the distributary canyons of Grand Coulee, except Dry Coulee, are thought to have functioned during the second flooding, the Adrian terrace is considered to be a part of the original fill and not of Wisconsin age, and all the deep canyons of Drumheller Channels are thought to date back to the earlier episode.

[3] Below an altitude of about 1,250, erratic bowlders occur on every formation and type of topography on the plateau. But these are a younger deposit (see *Journal of Geology*, Vol. XXVII [1919], pp. 489–506) and do not much overlap the scablands. Where overlap does occur, however, it is impossible to distinguish bowlders of the two categories by any difference in the amount of decay.

DEPTHS OF SNAKE AND COLUMBIA VALLEYS DURING THE EPOCH

Evidence on this question may be obtained at the debouchure of glacial drainage routes into these master valleys. Five of the nine such debouchures will be examined.

The rock floor of Moses Coulee is fully as low as that of Columbia Valley at the junction of the two. Both contain a great gravel fill here. Columbia River has cut through it, a depth of more than 300 feet. There is no such trenching in Lower Moses Coulee, but a well at Appledale penetrates 300 feet of this gravel without encountering bedrock.

The two cataracts of The Potholes and Frenchman Springs, which operated in the early part of the epoch, descended nearly the full height of the present Columbia Valley walls there. At The Potholes the glacial cataract can be traced down to less than 200 feet above the present Columbia (Fig. 7).

Koontz Coulee, 20 miles north of Pasco, is cut in the weak Ringold formation. It is 250 feet deep and a mile wide. It is floored with basaltic stream gravel from the scablands farther upstream. Though the Ringold silts extend down to the level of the Columbia at this place, the mouth of the glacial river channel hangs 200 feet above. No cataract could have been maintained here, as was done at The Potholes and Frenchman Springs, and the level of the Columbia of this epoch at this place is thus clearly recorded.

At the mouth of Palouse River, there are two parallel canyons in the scabland, one containing the river, the other dry. A basaltic butte separates them. It stands nearly in the center of the valley and its summit is between 350 and 400 feet above Snake River. It is a part of the original north wall of Snake River Valley, over which the gigantic cascade tumbled when the glacial flood broke across the preglacial divide from the north. This "Goat Island" testifies to the existence of a Snake River Valley at this place as deep then as now.

THE GLACIATION

Because the record of the ice sheet, from which came the streams that made the scablands, is best preserved on the basalt plain about the city of Spokane and along the south side of Spokane River,

this has been named the Spokane glaciation.[1] It assuredly is not an early phase of the Wisconsin glaciation. That is recorded by pronounced moraines on the plateau west of Grand Coulee, in Columbia Valley north of the mouth of Spokane River, and in Colville Valley north of Spokane River. The Spokane ice left no terminal moraine and very little ground moraine. The reverse relation exists regarding the glacial waters of the two glaciations, for the Spokane glacial waters were of prodigious quantity and the Wisconsin waters of little consequence. Furthermore, a long time elapsed between the two glaciations as shown by the relative volume of talus accumulations in tracts swept by glacial streams of the two epochs. Post-Spokane talus in almost all places stands three-fourths to four-fifths of the total height of the basalt cliffs, post-Wisconsin talus stands about halfway up on the cliffs of Grand Coulee.[2]

For the absence of a terminal moraine along the southern edge of the area reached by the Spokane ice sheet, the writer has as yet no satisfactory explanation. It seems clear, however, that a moraine never was deposited, rather than that it was once built and subsequently removed. The functioning of some scabland tracts absolutely required glacial ice against the north slopes of the unglaciated hills at their heads. Floated granite erratics among some of these hills also require blocking of valleys by glacial ice. Yet there is no evidence of lateral drainage along the ice front in contact with the unglaciated hills, to which might be ascribed the removal of a terminal moraine.

The Spokane glaciation cannot be dated very far back in the Pleistocene, else the scablands should have a soil mantle of eolian sand and dust and disintegrated basalt, and the hundreds of lakes in the old channels should have been destroyed.

Leverett has suggested that a pre-Spokane till beneath loess at Cheney is of Kansan age.[3] If it is, and if the post-Spokane glaciation is correctly ascribed to the Wisconsin epoch, the Spokane

[1] J H. Bretz, "Glacial Drainage on the Columbia Plateau," *Bulletin of the Geological Society of America*, Vol. XXXIV (1923), pp. 573–608.

[2] This question of differentiating Pleistocene epochs by talus accumulation will be discussed more fully in a separate paper.

[3] Frank Leverett, *Bulletin of the Geological Society of America*, Vol. XXVIII (1917), p. 143.

glaciation should be either Iowan or Illinoisan in age. Farther than this, the writer does not care to go. Ordinary criteria in use east of the Rocky Mountains for differentiation of drift sheets cannot safely be used for the correlation of these glaciations in Washington. The only one relied on here is the moraine-building habit of the Wisconsin ice sheet, a character which seems to have been world-wide.[1]

There were no channeled scablands on the Columbia Plateau before the Spokane glaciation. A mantle of loess, with a mature topography, completely covered it. The evidence for this conclusion is found in the great and remarkably persistent width of the Cheney-Hooper scabland tract throughout a length of 70 miles, and the various distributary courses out of it, some of which never were eroded to the basalt. These features never could have been formed, had spillways like those of the present existed. But with early escape southward retarded by the loessial hills and their small drainage ways, a wide spreading among them necessarily occurred, and some distributaries were able to cross to Crab Creek drainage.

A puzzling situation regarding glacial drainage exists in the vicinity of the small Spangle lobe. There is no adequate drainage route around it for glacial waters which came from Idaho and western Montana and entered the Cheney-Hooper scabland tract. Two spillways exist north of Mica, between Lake Spokane[2] east of this lobe and Pine Creek channel. Both have erratics in them, the highest at 2,550 feet A.T., but neither carried much water. There is no error involved in mapping this lobe because an ice dam at Spangle is required for the operation of the Pine Creek channel. This channel carried far more water than the Mica spillways, water derived directly from the Spangle lobe. Yet it also is inadequate for the drainage in question. And much more water went down the Cheney-Hooper scabland tract, in proportion to the immediately tributary ice edge, than passed through any other

[1] No till or other evidence of pre-Spokane glaciations has been found beneath the loess (save only the Cheney deposit). No till has been found in the scablands. The writer is unable to agree with J. T. Pardee who states (*Science*, Vol. LVI [December 15, 1922], pp. 680–87) that till occurs at "scores" of places on the plateau south of the limit of Spokane glaciation, as mapped in Plate III.

[2] Thomas Large, *Science*, Vol. LVI (September 22, 1922), pp. 335–36.

scabland tract except Grand Coulee in its later stages. It seems necessary, therefore, to assume a prominent *subglacial* drainage line, across the area covered by the Spangle lobe. This is best located along the preglacial valley of Lake (Marshall) Creek and the rock basin of Farrington (Fish) Lake. Out of this rock basin, just beneath the edge of the ice, the waters from the east emerged and joined those coming directly from the ice.

If the battle between the diluvialists and the glacialists, out of which has emerged our conception of Pleistocene continental glaciation, had been staged in the Pacific Northwest instead of the Atlantic Northeast, it seems likely that the surrender of the idea of a debacle might have been delayed a decade or so. Fully 3,000 square miles of the Columbia plateau were swept by the glacial flood, and the loess and silt cover removed.[1] More than 2,000 square miles of this area were left as bare, eroded, rock-cut channel floors, now the scablands, and nearly 1,000 square miles carry gravel deposits derived from the eroded basalt. It *was* a debacle which swept the Columbia Plateau.

[1] Except in the Hartline and Quincy structural depressions.

ALMANDITE AND ITS SIGNIFICANCE IN THE CONTACT ZONES OF THE GRENVILLE LIMESTONE[1]

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INTRODUCTION

The highly metamorphosed rocks of the Grenville series have attracted intense interest ever since Sir William Logan gave them a name and a place in geological literature. In view of the great wealth of geological information obtained from these old crystallines it would seem that they have nothing new to give to the science of geology and mineralogy, yet every time they are studied they yield some piece of information. The present article deals with one mineral, almandite, its origin, alteration, and instability. These changes are involved in the intense contact action produced by the great Laurentian Batholith.

LOCATION

The area which was studied previous to preparing this paper lies in Chatham Township, four miles northwest of the town of Lachute on the North Shore branch of the Canadian Pacific Railway 44 miles west of Montreal. The district is on the very edge of the Laurentian Highlands and the swift streams have cut moderately deep valleys for a short distance back from the broad fertile plain of the Ottawa Valley (Fig. 1).

GENERAL GEOLOGY

The rocks of Chatham Township are chiefly old crystallines, in part intensely metamorphosed sediments of early Precambrian

[1] The chemical and petrographical studies undertaken previous to preparing this paper were carried out at Columbia University under the guidance of Prof. J. F. Kemp, Dr. C. P. Berkey, and Mr. R. J. Colony to whom the writer is greatly indebted for assistance, advice, and helpful criticism.

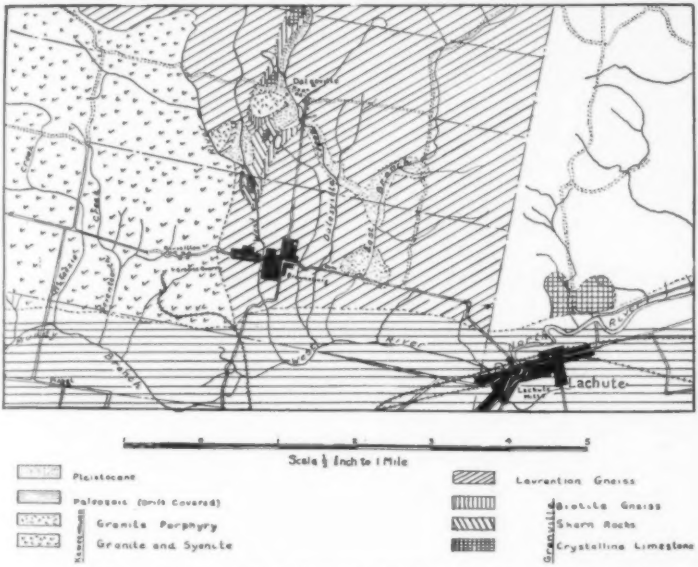


FIG. 1.—Map of the Grenville Series and related rocks of Chatham Township.

age, but mainly large intrusive batholiths of granite. Glacial débris of the Pleistocene ice sheet is strewn over the surface.

Pleistocene	Lake deposits
	Ground moraine
Keewenawan	Trap and microgranite
	Granite porphyry
	Granite and syenite
	Granodiorite
	Diabase
Laurentian	Granite
	Aegerite granite dykes
	Peridotite and pyroxenite
Grenville Series	Limestone

In this complex only the Grenville series and the Laurentian granite are of interest in this paper.

THE GRENVILLE LIMESTONE

The Grenville limestone is a blue, coarsely crystalline limestone composed of about 1 per cent of dolomite, 15 per cent of wollastonite

and diopside, and the remainder calcite. Minor amounts of phlogopite, garnet, chondrodite, scapolite, and sulphides are also found locally, but their occurrence is always such as to suggest introduction by magmatic emanations from the granite. An inclusion of blue limestone in an aegerite granite dyke of early Laurentian age gave the following analysis:

	Per Cent
Diopside and Wollastonite	16.4
Soluble Iron and Alumina	2.2
Calcite	81.0
Dolomite	1.1
	100.70

(The carbonates are given as dolomite and calcite rather than as dolomite or as calcite and magnesite, because when the polished surface of a specimen was treated with Lemberg's solution, it was found that the rock consisted of two definite carbonates, the one being calcite, and the other, in an amount demanded by the dolomite ratio, determined by a chemical analysis.)

Since this fragment of limestone was sealed off from all outside sources of material before the contact effects of the granite became pronounced, it is believed to represent very closely the composition of the original limestone; slight additions of silica from the dyke may have been received. On the other hand, since the dyke itself is somewhat pegmatitic, the limestone may have undergone mild contact-metamorphism before being effectually sealed away.

Away from the dyke the limestone has come under the influence of magmatic waters which have produced considerable silication. A typical analysis of the silicated limestone is as follows:

Proximate	Per Cent	Ultimate	Per Cent
Wollastonite and diopside	30.00	SiO_2	16.05
Soluble iron and alumina	0.25	Al_2O_3	0.14
Calcite	69.62	Fe_2O_3, FeO	1.54
Dolomite	0.15	CaO	46.65
		MgO	5.25
		Na_2O	.....
		K_2O	.....
		CO_2	30.80
			100.43

THE SKARN ROCKS

When this bed is followed along the strike for about 600 feet it changes to a rock composed of:

	Per Cent
Diopside, Wollastonite	47
Phlogopite	few flakes
Calcite, Dolomite	45
Pyrrhotite	2
Magnetite	1

After a short distance the bed comes under greater influence of solutions from the batholith and changes over to a skarn rock composed of:

	Per Cent
Orthoclase, Andesine	75
Almandite	10
Pyroxene	4
Calcite, Dolomite	8
Phlogopite	1
Magnetite, Pyrrhotite	2

At this point the bed passed beneath a thick cover of drift and the more advanced stages, which are of chief interest in this discussion, had to be studied in another locality where the succession was more complete. The locality chosen was at Campbell's Dam near Dalesville, where excavations had been carried on to some extent in preparing the base for a waterwheel. At this point the limestone was observed to change quite rapidly to a skarn rock containing larger garnets in greater abundance than in the former locality. Both the skarn rock and the garnets were analyzed and the results are reproduced below.

	Garnet Rock Per Cent	Garnet Per Cent	Garnet Per Cent
SiO_2	53.10	41.15	41.33
Al_2O_3	24.38	20.60	20.77
Fe_2O_3	2.86	Tr.	Tr.
FeO	11.21	30.09	30.05
MnO	Tr.	Tr.	Tr.
CaO	1.16	2.41	2.37
MgO	2.40	5.73	5.67
Na_2O	2.94	.....	.....
K_2O	1.49	.....	.....
H_2O	0.60	.....	.....
	100.05	99.98	100.19

In addition to garnet the rock contains numerous grains of magnetite and stringers of clear quartz. Feldspars are also abundant.

THE BIOTITE GNEISS

When the garnet zone is followed along the strike for any distance it passes into a biotite gneiss. Thin sections of this gneiss were examined and the rock was seen to contain about:

	Per Cent
Biotite	15
Quartz	25
Chlorite	15
Orthoclase	12
Andesine	15
Almandite	5
Sillimanite	3
Carbonates	3
Sericite	2

The garnet is heavily fractured and elongated in the plane of schistosity. Along these fractures it has altered to a green chlorite which itself is recrystallizing to form biotite. The sillimanite exhibits a change similar to that taking place in the garnet.

Both the biotite gneiss and the biotite of the gneiss were analyzed. The results of the analyses are given below:

	Biotite Per Cent	Biotite Gneiss Per Cent
SiO_2	36.93	55.91
Al_2O_3	22.32	18.44
Fe_2O_3	15.53	2.37
FeO	1.41	7.42
CaO	2.73	3.50
MgO	9.22	3.61
MnO	Tr.	Tr.
K_2O	3.07	1.13
Na_2O	3.12	2.08
CO_2, H_2O, F	4.68	4.74
	99.01	99.20

When the biotite gneiss is studied as a rock rather than as a petrographical curiosity, it is seen to be cut by innumerable pegmatite dykes which are connected with the Laurentian granite. These pegmatites have soaked the biotite-bearing rocks and it is almost inconceivable but that they must have added silica, a lumina, and alkalies to the original.

THE GRANITE GNEISS

To complete the discussion of the contact zones and the alteration of the almandite, it is necessary to include an analysis of the granite gneiss. Although this analysis is given in *The Superior Analyses of Igneous Rocks*, it is not typical of the Laurentian granite but rather of the granite gneiss, which is a complex or hybrid rock distinguishable from the granite by a higher percentage of ferro-magnesian minerals and a strong gneissic structure. The granite itself contains practically no ferromagnesian minerals and has at best a very poor streaked appearance when free from roof pendants of gneiss.

	Granite Gneiss* Per Cent
SiO_2	59.89
Al_2O_3	17.70
Fe_2O_3	1.95
FeO	2.71
CaO	2.53
MgO	1.56
MnO	....
K_2O	5.83
Na_2O	5.74
TiO_2	0.96
P_2O_5	0.17
CO_2	0.39
H_2O	0.29
	99.92

Analyst: Dittrich

* Annual Report of the C.G.S., Part O, 1899.

THE ZONAL ARRANGEMENT OF THE ROCK TYPES

The field relations show a zonal arrangement of rock types brought about by pegmatitic influences of the Laurentian granite on the Grenville limestone. The zonal arrangement is as follows:

Crystalline Limestone
|
Diopside, Wollastonite Skarn Rock
|
Feldspar. Garnet Rock
|
Biotite, Garnet Gneiss
|
Biotite Gneiss
|
Granite

In the field work, care was exercised in following the same bed, and where it was necessary to change from one bed to another the new one was picked up where it was almost identical with the one being studied. The changes outlined may be taken to represent the metamorphism of an average piece of Grenville limestone.

COMPOSITION CHANGES DURING METAMORPHISM

For the purpose of studying the metamorphism that has taken place, it is advantageous to recalculate all the ultimate analyses to molecular proportions. Molecular proportions are calculated by dividing the percentage weight by the molecular weight and multiplying by 1,000 to dispense with decimal points. The advantage of molecular proportions over percentage weights is that in chemical reactions, to form definite mineral compounds, a definite number of molecules of one oxide combine with a definite number of molecules of another. If the relative number of molecules of each oxide present is known, it is an easy matter to see the changes that have taken place. Although the results could not be recalculated with accuracy to mineral percentages on account of the mineral complexity, the method was deemed advisable in order to adhere more closely to the chemical principles in distinguishing recrystallization from replacement. Where simple minerals such as biotite and garnet are discussed, this method can be applied with obvious advantage.

Since the changes in the rock are indicative of the conditioning agencies, it is desirable to discuss them at some length before proceeding to discuss the changes in the garnet. The analyses of all the rocks recalculated to molecular proportions is given below:

	Limestone	Garnet Rock	Biotite Gneiss	Granite Gneiss
SiO_2	266	880	929	993
Al_2O_3	1	239	180	173
Fe_2O_3	9	18	15	12
FeO		156	103	38
CaO	832	21	62	45
MgO	137	60	90	39
Na_2O	0	47	34	62
K_2O	0	16	12	92

From the analyses it will be seen that the change is in the direction of equilibrium with the granite. For the purpose of calculating the additions and subtractions, some constituent of the

original rock must be considered to remain behind throughout all the changes. Alumina seems to fill this place to perfection, since once it is precipitated it seems to migrate again only under extreme conditions, such as resolution in a batholithic magma and discharge in pegmatites. Some additions of alumina may occur but their effect will be pointed out later.

FIG. 2.—Highly silicated limestone. (Magnification about 30 diameters. Ordinary light.) The photograph shows the initial development of metapoikilitic structures with increasing amounts of diopside and wollastonite (strong cleavage). The cloudy area near the center of the field represents a pseudoeutectic intergrowth of dolomite and calcite. Calcite forms the remainder of the field.

Consider first the change from limestone to garnet rock (Figs. 2 and 3). Then since the garnet rock is formed from the limestone, replacement of the latter must continue until there are 239 times as many molecules of alumina as before. If the garnet rock was formed by simple recrystallization of the limestone, the difference expressed in the difference column would remain unadjusted and the excess material would then have to be removed in solution and would produce a great decrease in volume. But all field evidence points to practically no change in volume in this rock.

		Limestone $\times \frac{239}{1}$ =	Garnet Rock	Difference
SiO_2.....	266	=63574	880	−62694
Al_2O_3.....	1	239	239	0
Fe_2O_3.....	9	2151	18	−1977
FeO.....			156	
CaO.....	832	198848	21	−198827
MgO.....	137	32743	60	−32683
Na_2O.....	0	0	47	47
K_2O.....	0	0	16	16

The change from limestone to garnet rock involves a loss of everything but alkalies. Much of the lime and magnesia has been removed in solution and carried into the upper zones which are now

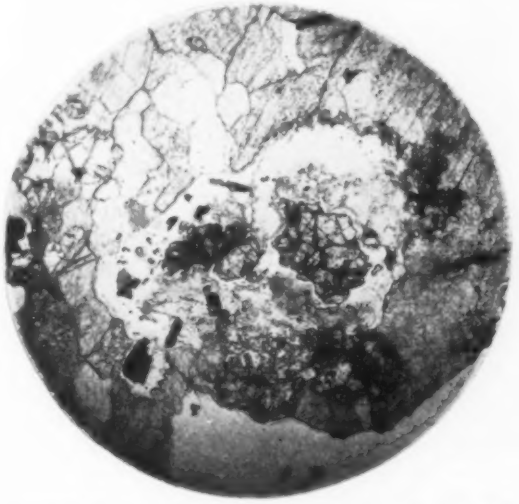


FIG. 3.—Garnet in limestone. (Magnification about 30 diameters. Ordinary light.) The photograph shows garnet (high relief) in the center of the field; it is surrounded by a broad kelyphite rim, which seems to be an incipient stage in its formation. The remainder of the field is largely carbonates.

eroded away. The apparent loss of iron and silica is due to replacement of the lime and magnesia by these constituents. From chemical analyses alone it is difficult to follow the changes, since an addition in one rock has the same effect as the loss of the same

constituent in another. Relative changes only can be determined from the composition, so that in the following interpretations the direction of the changes is based upon microscopic study and the quantitative relationships obtained from the chemical analyses.

In the garnet zone the structure indicates that lime and magnesia went into solution and were replaced by silica, alumina, and iron

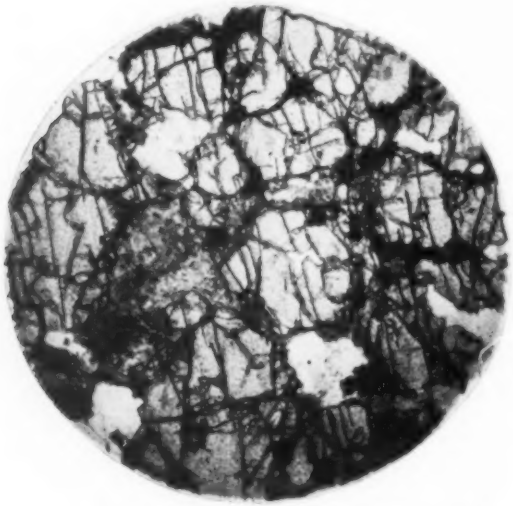
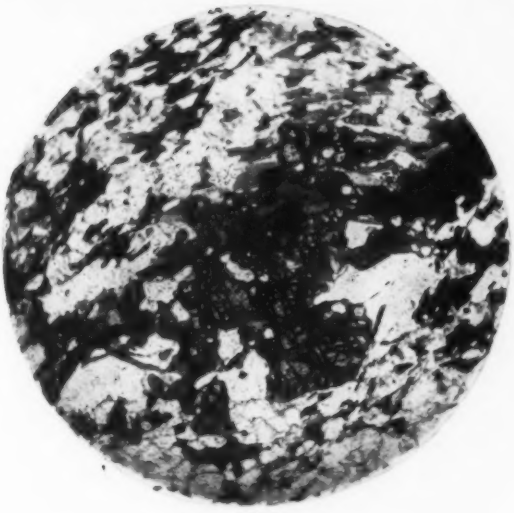


FIG. 4.—Garnet from garnet feldspar skarn rock. (Magnification about 30 diameters. Ordinary light.) The photograph shows the garnets traversed by two sets of fractures, the one more prominent than the other. The clouded areas are feldspar and the clear colorless areas quartz. The black opaque spots are magnetite.

in the form of garnet and feldspar. Selective solution and precipitation have doubtless played an important part, but the exact amount of each cannot be determined in as close a manner as in the succeeding zones. The high differences simply indicate almost complete replacement as being the dominant process in this zone as compared with dominant recrystallization in those following (Fig. 4).

Consider the change from garnet rock to biotite gneiss (Fig. 5, *A* and *B*). In this case the change is not so great; replacement



A

B

FIG. 5, *A* and *B*.—Biotite garnet gneiss. (Magnification about 30 diameters. Ordinary light.) The photograph (*A*) shows the garnets traversed by two sets of fractures one of which is parallel to the schistosity. The dark areas along the fractures are chiefly chlorite. The lighter colored dark areas, the location of which is indicated on the diagram (*B*), some of which are very minute and parallel to the schistosity, are biotite.

doubtless plays an important part, but the dominant processes are granulation and recrystallization with increase in rock volume as shown by the change of garnet and sillimanite to chlorite (Fig. 6).

FIG. 6.—Biotite gneiss. (Magnification about 30 diameters. Ordinary light.) The photograph shows a great abundance of biotite (dark) and a complete absence of garnet due to the completion of the alteration. A single elongated prism of sillimanite occurs in the center of the field. Along the cleavages it is altered to chlorite (dark). The rectangular outline at the end of the prism is a basal section of sillimanite altered to chlorite.

As before, the alumina of the original rock is assumed to remain behind and additions and subtractions of the other constituents to occur to produce the required changes.

	Garnet Rock	= Biotite Gneiss	$\times \frac{2.19}{1.80}$	= Difference
SiO_2	880	929	= 1235	= 355
Al_2O_3	239	180	239	0
Fe_2O_3	18	15	20	2
FeO	156	103	137	−19
CaO	21	62	82	61
MgO	60	90	120	60
Na_2O	47	34	45	−2
K_2O	16	12	16	0

The change involves a gain in silica, lime, and magnesia and a loss of ferrous iron and soda. In this case the change in each constituent, with the exception of silica, is small. If complete replacement had occurred, the difference ought to have been infinity and, by analogy, rocks involving smaller differences in composition ought to be regarded as due to recrystallization of the minerals of the preceding zone with smaller changes brought about by replacement and addition.

Now since there has been an increase in lime and magnesia in the biotite gneiss, it must have been supplied either by the next zone nearer to the batholith or allowed to pass through that zone in solution because it did not require so much magnesia and lime to maintain a state of equilibrium. Consequently the granite gneiss ought to show an impoverishment of these constituents and the silica might also show a decrease, although that would not necessarily follow because of the intense pegmatitization. Suppose the composition changes of granite gneiss are examined.

	Biotite Gneiss	Granite Gneiss	$\times\frac{180}{173}$	Difference
SiO_2	929	993	= 1033	104
Al_2O_3	180	173	180	0
Fe_2O_3	15	12	12.5	−2.5
FeO	103	38	40	−65
CaO	62	45	47	−15
MgO	90	39	41	−49
Na_2O	34	62	65	31
K_2O	12	92	96	84

The granite gneiss is highly impoverished in lime and magnesia, while the additions of silica have decreased somewhat over the amount being introduced into the preceding zone.

Throughout the calculations the alumina has been assumed to remain constant in the original rock and the changes to occur by replacement of the other components. Such condition may be seriously questioned since the volume of the biotite gneisses has been increased somewhat over 15 per cent by the direct injection of pegmatite dykes. A corresponding increase might also be expected to occur through simple soaking of the rocks by pegmatitic juices. When the analyses, however, are recalculated on the assumption of an increase of from 15 per cent to 20 per cent in alumina, the only change is the establishment of equilibrium of the iron oxides;

that is, the amount of iron oxide remains constant in the rock. The alumina however accumulates above an amount required to maintain an equilibrium, and when this condition is attained, some alumina and iron oxides are held by the pegmatitic solutions, which transport them to a zone where lack of equilibrium exists. This happens to occur where the stresses in the rock are such as to permit the formation of garnet and feldspar, that is the stress on the crystal is essentially static load. The available iron and alumina combine with sufficient silica to form almandite. Magnesia and lime also enter into the combination to form grossularite and pyrope molecules. Some of the remaining available alumina combines with the alkalies and lime to form feldspars while a small amount is carried into the limestones to form mica and other aluminous minerals.

THE DESTRUCTION OF THE GARNET ZONE

As soon as the garnet feldspar zone has formed, the batholith is free to extend its activity farther into the limestone. Lateral stresses caused by mountain folding, igneous activity, and increase in rock volume produced fracturing in the garnet which then became overwhelmed in a zone in which it was unstable. Stress from one direction causes the almandite to break down into chlorite when in the presence of the circulating magmatic waters. The chlorite readily recrystallizes to biotite under the influence of these alkaline solutions.

For the purposes of studying the alteration of the garnet, the composition of both the mica and the garnet have been recalculated to molecular proportions. (In the case of the garnet it will be noticed that the material used was not entirely freed from the quartz stringers which cut the almandite and which in ordinary light look very much like it.)

	Garnet	Biotite
SiO_2	685	613
Al_2O_3	201.5	219.5
Fe_2O_3	...	14.8
FeO	419	19.6
CaO	42.5	48.7
MgO	142	228.5
Na_2O	...	50.3
K_2O	...	39.3

The change in composition shows quite clearly the effects of magnesia bearing alkaline solutions. Apparently these found easy channels along the fracture planes of the garnet where alteration is most intense.

If the garnet is corrected for the amount of quartz, which appears to be high in the molecular proportions but is really very small, the molecular composition would be:

	Garnet	Biotite	Difference
SiO_2	603	613	10*
Al_2O_3	201.5	219.5	18*
Fe_2O_3	...	14.8	385
FeO	419	19.6	
CaO	42	48.7	5.7*
MgO	142	228.5	86.5
Na_2O	...	50.3	50.3
K_2O	...	39.3	39.3
H_2O / F	...	292.0	292.0

* These differences are so small that they might readily arise through errors in sampling and in analysis as well as other variations which cannot be guarded against. For these reasons they cannot be regarded as a part of the changes during the reaction.

The F_2O_3 and FeO of the biotite are grouped together, the only significance of the ferric iron being that it indicates a very slight oxidation which, although showing in the mineral, does not occur so prominently in the rock. The outstanding differences then are the replacement of the ferrous iron by MgO, Na_2O, K_2O, H_2O, and F. Although replacement has increased the number of molecules by almost exactly one-third, which is a remarkable feature, it does not seem that the reaction could be due to simple reorganization. All of the variables are divalent and with the exception of hydroxyl and fluorine are chemically positive elements. The definite 2–3 ratio between the positive elements, alkalies and magnesia, and the negative elements, hydroxyl and fluorine suggests that a rather definite hydrated chemical compound had replaced the FeO during this reaction. If such is the case, the compound was one of the solutes in the highly heated emanations from the granite magma. Whether the fluorine and water existed separate at the existing temperature and pressure is unknown, but the oxidation of the

ferrous iron in the garnet would seem to indicate that some reaction similar to that given below had taken place:

$$2(H_2O+2F)=4HF+O_2$$

:—Nascent oxygen.

$$4FeO+O_2=2Fe_2O_3$$

From garnet-: :—In biotite containing F.

Some reaction giving this final result does take place, and since the entire oxidation that is transpiring in the rock accompanies this change which involves the fluorine, it would seem that the above condition represents the instability in this zone. If this is correct the reaction represents the critical point above which water and fluorine can exist together as definite and distinct components.

The increase in volume caused by this reaction (the change of almandite to biotite) must have developed a pressure normal to the fractures in the garnet so that where the chlorite recrystallized to biotite the mica flakes are elongated along a certain plane to produce schistosity in the rock. Two directions of fracture are present in the garnet, one parallel to the schistosity and another at right angles to it. The chlorite along the fractures parallel to the schistosity seems to have recrystallized more readily than along the fractures at right angles to it. Fracturing in the garnet seems to have been caused by folding of the Grenville series; thus the planes of schistosity and the banding of the gneisses produced by lit-par-lit injection are controlled by the direction of the axes of the folds of the Grenville series. Migration of the biotite along the planes of schistosity, for short distances, has taken place in some cases, owing to the strong tendency of the biotite to crystallize along these planes.

CONCLUSIONS

The Grenville crystalline limestones have been subjected to intense contact action by the intrusion of the Laurentian granite. Garnet zones were developed in the limestones.

The garnet is of an unusual type containing almandite, grossularite, and pyrope in the molecular ratio 20–2–7. Kemp has collected as complete a list as possible of the analyses of garnets

found in contact zones previous to 1912, and has shown that the usual range is nearly limited to andradite or grossularite.

TABULATION OF ANALYSES OF GARNETS FROM CONTACT ZONES TO SHOW NEW TYPE FOUND IN SILICATED LIMESTONES AROUND THE LAURENTIAN GRANITE

GARNETS FROM CONTACT ZONES OF POST-CAMBRIAN INTRUSIVES*

No.	Percentage Proportions					Molecular Proportions				
	Andra.	Gross.	Spess.	Alman.	Pyro.	Andra.	Gross.	Spess.	Alman.	Pyro.
1	92.6	3.00	0.86	0.48	0.43	1825	66	19	8	11
2	89.52	.....	.....	.....	.....	1765	.....	.....	.....	.....
3	88.74	.....	.....	.....	.....	1749	.....	.....	.....	.....
4	87.57	1.60	0.63	1.46	2.32	1725	35	14	24	57
5	87.54	3.26	1.01	0.90	.....	1725	72	23	15	.....
6	86.36	2.96	0.79	.....	.....	1700	66	18	.....	.....
7	82.15	4.08	.....	.....	5.40	1620	90	.....	.....	134
7a	76.83	8.80	.....	.....	.....	1512	195	.....	.....	.....
8	67.62	6.02	0.91	.....	.....	1332	133	20	.....	.....
9	64.62	22.44	.....	.....	3.64	1272	496	.....	.....	90
10	63.50	13.50	1.38	14.35	.....	1250	299	31	240	.....
11	61.10	31.00	.....	.....	.....	1201	687	.....	.....	.....
12	59.04	15.20	.....	.....	.....	1165	337	.....	.....	.....
13	58.52	9.00	14.40	.....	.....	1152	199	322	.....	.....
14	53.80	27.90	.....	.....	.....	1060	618	.....	.....	.....
15	53.40	37.40	1.20	3.50	4.40	1052	829	27	58	109
16	52.80	23.20	.....	.....	.....	1040	515	.....	.....	.....
17	45.15	41.16	.....	.....	.....	890	911	.....	.....	.....
18	44.16	47.82	0.68	2.99	1.31	870	1060	15	50	32
19	35.46	53.10	.....	.....	.....	699	1179	.....	.....	.....
20	24.90	42.00	.....	.....	.....	490	931	.....	.....	.....
21	24.28	40.80	.....	.....	.....	479	905	.....	.....	.....
22	23.50	48.40	.....	.....	.....	463	1071	.....	.....	.....
23	23.12	65.72	.....	.....	.....	455	1455	.....	.....	.....
24	21.13	69.26	.....	1.61	1.44	416	1535	.....	27	36
25	9.30	60.80	.....	.....	.....	183	1347	.....	.....	.....
26	1.92	55.08	.....	.....	.....	38	1221	.....	.....	.....

* *Trans. Can. Inst. of Min. and Met.*, XV., p. 178.

TWO GARNETS FROM THE CONTACT ZONES OF PRE-CAMBRIAN INTRUSIVES†

No.	Percentage Proportions					Molecular Proportions				
	Andra.	Gross.	Spess.	Alman.	Pyro.	Andra.	Gross.	Spess.	Alman.	Pyro.
1	.....	5.82	Tr.	76.75	17.43	.....	129	Tr.	1283	435
2	.....	5.82	Tr	76.75	17.43	.....	129	Tr.	1283	435

† This paper.

Subsequent changes in crustal and chemical stability and advance of the contact zones during the final stages of consolidation

of the granite have brought about fracturing in the garnet and its alteration to biotite mica. In the mica much of the lime is retained and magnesia increases; both soda and potash have been added.

In the garnet rock fractures traverse the equidimensional almandite grains. Folding has extended the fractures and elongated the grains; magmatic waters have brought about their alteration to chlorite. Recrystallization of the chlorite in the presence of alkaline waters containing fluorine had produced biotite. The increase in volume has aided the lateral thrust of mountain folding and developed a schistosity in the biotite gneiss.

In the contact zones the rule seems to be "Each succeeding zone is enriched in those constituents in which the preceding one is impoverished. There is no indication of the reverse being true; that is, no reverse reaction is taking place." The changes after the formation of a zone are due to the extension of the contact effects farther into the limestone and a burial of the old zone in a new one.

Pegmatitization and replacement were the agents that changed the rock. Replacement is important in the garnet zones, recrystallization in the biotite gneiss zone, and pegmatitization in the granite gneiss zone. All the processes more or less overlap.

If metamorphic rocks can be spoken of in terms of youth, maturity, and old age, the crystalline limestone might be considered as representing youth, the skarn rocks maturity, and the granite gneiss the extreme old age of the Grenville limestone.

A NOTE ON JOINTAGE AND THE APPLICATION OF THE STRAIN ELLIPSOID

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INTRODUCTION

In conducting a class in field geology across a portion of the superb section exposed in the walls of Merced Canyon, just below the Yosemite Valley, the writer recently observed an interesting case of jointage which appears to deserve note.

The section at its upper end includes the massive exposures of the granitic rock of the Sierra batholith, such as in the sheer wall of El Capitan and adjacent exposures, and it extends down the canyon of the Merced across the western margin of the batholith and into the adjoining belt of metamorphic rocks.

DISTRIBUTION OF THE JOINTS

The jointage here considered is best developed and most conspicuous on the immediate margin of the plutonic mass, though jointing is by no means restricted to this portion of the section.

In the massive walls of the Yosemite Valley strong and persistent joints are to be seen. Some are very steep, some inclined at moderate angles, still others are at low angles. No attempt was made to study these in detail, but one gets the impression from the floor of the valley that the moderately inclined joints, dipping both easterly and westerly, are somewhat more conspicuous than the others. In this upper portion of the section, some 10 miles from the western margin of the batholith, the joints are as a rule widely spaced and the rock loses none of its massive character as a result of the fracturing.

Here the average rock type is a medium-grained light-colored granodiorite. As one proceeds down the canyon, toward the outer portion of the granitic mass, changes in both composition and

structure are noted. Various types of biotitic and hornblendic schlieren become more and more abundant. These include rounded and orbicular and more angular types, as well as irregular areas and dike-like schlieren. Pegmatite and aplite dikes become more numerous, and more joints appear. Finally the intrusive loses its massive character entirely in a marginal zone characterized by the extreme fracturing shown in the accompanying photographs.

The outcrops here pictured are located in the granodiorite, about 100 feet inside the contact with the metamorphic rocks of the "Bedrock Series," in the railway cut one mile below El Portal. Jointage of similar character continues into the metamorphic rocks of the contact zone, but in these heterogeneous rocks it is not so regular and the analysis of the joints becomes more difficult.

CHARACTER OF THE JOINTS

As is evident from the photographs, the exposures are here cut by three dominant sets of joints. These are intersected by numerous less regular and less persistent fractures of variable extent. What we may here consider as the master joints include the nearly vertical set of dike joints and the two sets of moderately inclined joints.

The steep joints comprising the first set are often made conspicuous by the presence along the joint plane of thin sheets of white aplite, varying from less than one-half inch in thickness up to six inches. These aplite dikes are here practically restricted to the system of steep joints, though here and there a dike may be seen to break away from the main joint plane and follow a minor, slightly divergent, subsidiary joint, as is well shown in the outcrop of Figure 1. The strike of the joint planes of these dike joints at the locality studied is about N. 85° W.; the dip varies 5° either way from the vertical, though most frequently it is about 85° to the south.

The two sets of inclined joints are somewhat stronger and more regular than the dike joints. In the locality studied one of these sets strikes N. 50° W. and dips about 35° to the south; the other set strikes N. 15° W. and dips about 20° to the east. Though not spaced with exact uniformity the effect of these intersecting fracture

systems in the plane of the section is to divide the granitic rock quite regularly into large and small rhomboidal joint blocks, crossed by the less conspicuous fractures of the steep joint system, as illustrated in Figures 1, 2, and 3.

FIG. 1.—Jointage in granodiorite. Looking north. The steep joint system is indicated by aplite dikes. The heavy inclined joints dip east and are intersected by a third system dipping to the left, only faintly seen in this exposure.

Very frequently the steep joints with their accompanying dikes are slightly offset along the fractures of the east-dipping inclined set, the strongest set of the three, so that this set is really a series of small thrusts with minute throw, as shown in Figures 2 and 3.

The thrust character of the strong joint plane dipping to the right in Figure 3 is shown in the offset of the joint and dike indicated by the pick. The displacement along the thrust plane is here 20

FIG. 2.—Joints in granodiorite. Looking north. Rhomboidal joint blocks produced by the intersection of the two sets of inclined joints, crossed by the fainter steep dike joints. The strong joints dipping to the right are frequently small thrust fault planes, as in the one at the bottom of the view.

FIG. 3.—Thrust character of many of the east-dipping joint planes, indicated by the displacement of the dike near the pick.

inches. A small subsidiary block on the extreme right, having evidently acted somewhat as a unit, is seen to be in turn crossed by short joints of the two inclined sets.

RELATIVE AGE OF THE JOINTS

It is evident from the facts related above that the fractures of the three systems were not all developed at the same time. Those of the steep set are clearly the oldest. They constituted favorable planes for the injection of the aplitic magma. Still later they were broken across by the development of the joints of the inclined sets, and they were displaced upward and westward along the planes of the east-dipping set, a set which became a series of minute thrusts, outward from the batholithic mass.

APPLICATION OF THE STRAIN ELLIPSOID

There would appear to be good reason to accept the two intersecting sets of inclined joints as an illustration of failure by fracture under lateral compression. It is obvious that, as the result of the summation of the minute thrusts, the rock mass has been elongated upward, in the direction of easiest relief, with shortening in approximately the horizontal direction.

Assuming that the two inclined sets of joints are closely related in origin we may apply the strain ellipse in the customary manner to the plane of the outcrops of the photographs. It is obvious that the major axis, representing the direction of elongation of the rock mass, should be in approximately the vertical direction. The planes of maximum shear are evidently represented by the two inclined sets of joint planes, as in the adjoining sketch (Fig. 4). Here, then, the major axis of the ellipse is the bisectrix of the *obtuse* angle made by the traces of the shear joints in the plane of the outcrop.

The fact that we are here left in no doubt as to the correct orientation of the ellipse makes the analysis of the joints of these outcrops especially desirable. Exactly similar jointage is often encountered in folded sedimentary rocks. In the interpretation of concealed structure from the study of such joints it is often difficult to determine what is the correct position of the ellipse.

Unless some good indication of the direction of movement of the beds in folding is found one may be left in doubt as to the interpretation.

In the present case slipping has obviously occurred on one of the sets, up and to the left. Let us suppose that, instead of granodiorite, these outcrops consisted of massive jointed sandstone or quartzite, in the vicinity of an axis of folding, and that the east-dipping joints were bedding joints. Since the movement on these has been upward and to the left the customary inference would be that the outcrops belong to the east (right) limb of a normal anti-

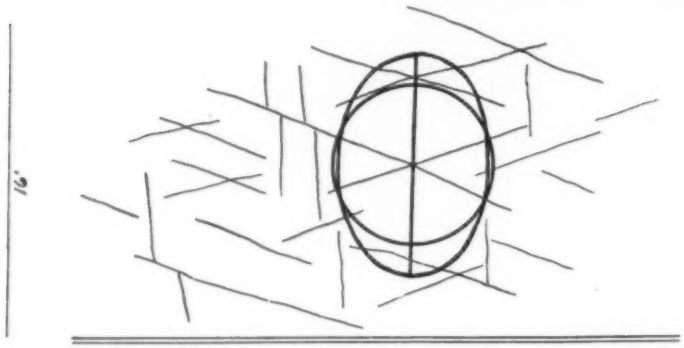

FIG. 4.—Position of the strain ellipse when applied to the outcrop shown in Figure 2. The acute angle faces the direction of shortening. This does not illustrate a simple distortion, as there is an increase in volume.

cline. In this supposition the second set of joints crosses the bedding joints obliquely, as a set of inclined compression strike joints facing the direction of movement with an *obtuse* angle.

Now, if the direction of movement on the bedding joints were not known, the student might be tempted to prophesy the position of the anticlinal axis upon the erroneous assumption that the acute angle between inclined joints of this character points in the direction of movement, for this is indeed the inference one gets from the customary diagrams of texts showing the strain ellipse applied to joints. If such an application be made to the case here considered we reach the surprisingly erroneous conclusion that the outcrops belong to the right limb of an overturned syncline.

The danger of thus misinterpreting joints is obviated only if the student disregards the customary sketches and remembers that the acute bisectrix may or may not be the major axis of the ellipse; that the angle between bedding joints and inclined compression joints facing the direction of movement may be acute, or obtuse, or 90°, depending upon the brittleness of the rock at the time of the development of the joints. In the case under study it is the obtuse angle that faces the direction of movement.

The importance of these considerations has recently been re-emphasized by Bucher, calling attention to the well-known experimental results embodied in Hartmann's Law that "in brittle materials, the acute angle formed by the shearing planes is bisected by the axis of maximum compression, and the obtuse angle by the axis of minimum compression which is generally negative, representing tension" in agreement with the mathematical results of Mohr.[1] It is also shown, from the consideration of Karman's experiments, that the angle between the shearing planes becomes less and less acute with decrease in brittleness, the usual models and diagrams giving expression "only to that case where the planes of shearing form an obtuse angle in the direction of compressive stress."[2]

The exposures here discussed are, then, a field illustration of the case where the acute angle lies in the direction of compressive stress, which, as suggested by Bucher, is the characteristic angle for brittle materials. The occurrence represents Bucher's third case of the application of Hartmann's Law, where the greatest compression is approximately horizontal.[3] It is evidently similar in many respects to the case of the symmetrical shearing plane faults in the limestone of the Hamilton Shale, referred to by Bucher and photographed and described by Sheldon.[4] Here likewise the acute bisectrix is the axis of maximum compression.

[1] Walter H. Bucher. "The Mechanical Interpretation of Joints," *Jour. Geol.*, Vol. XXVIII (1920), p. 712.

[2] *Ibid.*, Vol. XXIX, pp. 13–14.

[3] *Ibid.*, Vol. XXVIII, p. 727.

[4] *Ibid.*, p. 728. Also Pearl Sheldon, "Some Observations and Experiments on Joint Planes," *Jour. Geol.*, Vol. XX (1912), pp. 60–61.

THE GLACIAL GEOLOGY OF GRAND MESA, COLORADO

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The great number of lakes reported on and about Grand Mesa, Mesa County, Colorado, has long suggested glaciation to the writer, notwithstanding the fact that the mesa, which includes several townships, rises but little over 11,000 feet above sea-level and does not extend above timber line. However, I have never found any reference to glaciation in the literature of the region. In company with Mr. John P. Byram, of Mesa, Colorado, two weeks were spent on the mesa in August, 1923, with saddle and pack horses, accompanied for one day by Mr. Erwine Stewart, also of Mesa, who had told us of phenomena indicating glaciation. The evidence of intense glaciation is everywhere abundant around the north and east sides, and on a considerable part of the top, presenting two distinct types that I have not yet seen elsewhere in Colorado.

The ancient glaciers of Colorado, though confined to the mountainous portion, covered, at their greatest expansion, over 10 per cent of the total area of the state. Most of them originated along the crests of the higher mountain ranges above the 11,000 foot contour, and extended down deep, pre-existent gulches as typical mountain-valley glaciers ranging from three to sixty miles in length. Definite, well-marked cirques occupy the heads of these gulches, well above the present timber line. But in contrast with the common type of glaciation in the Rocky Mountains, deep gulches are absent from Grand Mesa, and cirques, if present at all, are so vague and ill-defined as to be scarcely recognizable.

Grand Mesa is bounded on all sides by a precipitous escarpment of basalt, a very important factor in the glacial geology of the region. The highest point determined by us by aneroid barometer

reading, near the eastern end, is only 11,300 feet, though that was along the trail and therefore not quite the highest point of the mesa. The top is a slightly rolling plateau, sloping gently to the southwest. At the base of the escarpment is a shelf two or three miles wide, extending all along the north and east sides, covered by an intricate network of moraines, with scores of lakes and ponds, mostly morainal. Near the escarpment, which provided an abundant supply of detrital material, the moraines are mostly high, narrow, steep sided, and strike one another at various angles, in many places forming basins of small radius. The more distant moraines, representing the greatest extension of the ice, perhaps two or three miles from the cliff, are lower and not so strongly defined.

Probably the less pronounced character of the more distant moraines is partly due to the fact that they have been longer uncovered and subjected to greater erosion, and partly to their greater distance from the source of supply. It is likely also that at the time of the greatest extension of the ice the cliff was more completely covered with ice and snow, so that less material fell upon the surface of the glacier to be carried down to the moraine.

The number, size, character, and relation to one another of the moraines near the cliff record frequent and rapid changes in the shape of the ice-front, with numerous small ice tongues extending beyond the general line of the front. This would be expected from the fact that the glacier (or glaciers) was doubtless formed and maintained chiefly by drifted snow, driven over the top by prevailing westerly winds and lodging along the rim of the cliff. Even now huge drifts remain along the rim throughout the summer. The drifts must have varied more or less from season to season, from cycle to cycle, and as a result the points of maximum and minimum accumulation, and consequently of greatest extension and contraction, must have shifted from time to time.

It seems likely, from our rather hurried investigation, that at the time of greatest expansion the ice along the north and east sides constituted one continuous glacier from ten to fifteen miles, perhaps more, in width and not more than three or four miles long. Such a condition would be unfavorable to the formation of well-defined cirques, though in its inception probably a number

of small glaciers formed and later coalesced, and in its retreat it was doubtless broken up into a number of separate ice masses before final disappearance, which history would be reflected in the number and character of the moraines.

The glaciated shelf is everywhere covered with a thick sheet of ground moraine. For this reason there is a great scarcity of *roches moutonnées* and glacial striae.

On top of the mesa different conditions produced very different results. There was no crest or cliff back of which drifting snow could accumulate for the formation of a glacier which would extend to the southwest, or that would furnish abundant material for the construction of moraines. A large, broad glacier, or perhaps more than one, formed on or near the highest point along the northeast rim and moved in a southwesterly direction along the gentle slope of the mesa. It built a long, low terminal moraine across the mesa near its middle, which now incloses several shallow lakes. Owing to the absence of ground moraine, its course is marked by numerous low, well-rounded *roches moutonnées*, but recessional moraines are absent or scarce and not well defined. The basalt weathers readily, so no traces of glacial striae were observed, but fluting was observed in several places. There are large morainal lakes on top near the eastern end of the mesa, but time did not permit their investigation. We found no clear evidence of glaciation on the western half of the mesa top. Because of the topography we believe the glacier (or glaciers) on top could not have been formed to any large extent by drifted snow, and that the glacier was probably rather shallow in proportion to its width and length.

EDITORIAL

With our rapidly growing science the volume of geologic literature has increased so enormously in recent years that it is no longer possible for any single geologist to keep up with the advances made in all the different branches of the subject. Instead, the active worker is now forced to pass hastily over many important papers which do not bear immediately on his own chosen field and to concentrate his energies chiefly upon those more immediately affecting his own specialties. A growing need has been felt and expressed for brief abstracts which, by pointing out the essentials of published papers, will enable each reader to determine quickly the value to him of any article. In response to this need, the *Journal of Geology* will inaugurate the policy, commencing with the first number of 1924, of printing an abstract in small type at the beginning of each article. Contributors are therefore kindly requested to include in each manuscript such an abstract, which should not exceed 250 words in length. This abstract will replace the table of contents, so often used in the past, but it need not necessarily exclude the use of a supplementary summary of conclusions at the end of an article if that seems also advisable. It is hoped that this new policy will make the *Journal* more useful to its readers.

R. T. C.

REVIEWS

Chief Results of a Preliminary Analysis of the Earth's Magnetic Field for 1922. By LOUIS A. BAUER. Terrestrial Magnetism and Atmospheric Electricity, Vol. XXVIII, Nos. 1 and 2 (March–June, 1923), pp. 1–28.

SUMMARY OF CHIEF CONCLUSIONS

"27. The chief conclusions reached in the present paper are as follows:

"*a*. It is necessary to recognize that the Earth's total magnetic field at any one time is apparently composed, to the extent of about 94 per cent, of an internal magnetic potential system, *I*, and to the extent of about 6 per cent, of an external magnetic potential system, *E*, plus a non-potential system, *N*. . . . .

"*b*. Instead of the three systems, *I*, *E*, *N*, the exact evaluation of each of which rests somewhat on assumption as to the character of the causes producing the observed magnetic field of the Earth, we may say that in order to represent satisfactorily the observed magnetic data, it is necessary to recognize the existence of three distinct systems, namely, the *X*, *Y*, *Z* systems. The *X*-system is responsible for the total component acting on the magnetic needle in the direction positive towards geographic North; the *Y*-system, for the total component in the direction positive towards geographic East; and the *Z*-system, for the total component acting in the vertical direction positive towards the nadir.

"*c*. It must be recognized that the magnetic secular-variation system is as complex as the Earth's total magnetic field existing at any one time, and that in addition to changes in the direction of magnetization with the lapse of time there is also a change in the average equivalent intensity of magnetization. The magnetic axis of one component of the total secular-variation system may show a reverse motion in latitude and longitude to that shown by another component. On the whole, as a resultant effect, it would seem that the north end of the axis of the Earth's internal magnetic field during the past 80 years has been moving slowly towards the west, and apparently at the same time slowly towards the equator. It is not possible to speak at present of any period of complete

revolution of the magnetic axis about the Earth's axis of rotation. (Were the average annual rate of motion of the magnetic axis, as indicated during the past 80 years, to continue unaltered, a complete revolution would require about fifteen thousand years, or more. Shorter periods deduced from analyses in which no account is taken of possible change in intensity of magnetization, besides change in direction of magnetization, or obtained from a discussion of the secular change in limited regions of the globe, evidently cannot be regarded as pertaining to the Earth's field as a whole.)

"*d*. The average equivalent intensity of magnetization of the Earth has been steadily diminishing during the past 80 years at the average annual rate of about 1/1500 part. How long it will continue to diminish, whether there will be a time of no change, or whether it will later increase, are questions that cannot be answered at the present time. (Whether the annual rate of loss is actually variable, as would apparently be indicated on the basis of the magnetic charts of 1885 and 1840–1845, is another matter which must await completion of additional analyses to be made for epochs for which magnetic data of known reliability are available.)

"*e*. A suggestive effect, dependent apparently upon the distribution of land and water, has been disclosed, namely, that the average equivalent intensity of magnetization for corresponding parallels north and south, is generally larger for the land-predominating parallel than for the ocean-predominating parallel.

"*f*. For 1922 we have for the Earth's uniform internal magnetic field, as deduced from magnetic observations over 86 per cent of the Earth's surface (60° N. to 60° S.), the following data, which it is expected will be only slightly modified by the final analysis, R being the Earth's mean radius (6.37×10^8 cm.):

M_p = Component of magnetic moment parallel to axis of rotation
$= 0.3047\ R^3 = 7.88 \times 10^{25}$ C.G.S.
M_e = Equatorial component of magnetic moment
$= 0.0618\ R^3 = 1.60 \times 10^{25}$ C.G.S.
M = Resultant magnetic moment[1] $= 0.3109 R^3 = 8.04 \times 10^{25}$ C.G.S.
$M_p = 4.93\ M_e$.

"If the Earth's magnetism were distributed uniformly throughout its volume, as it probably is not, the average intensity of magnetization would be 0.074 C.G.S. The magnetic axis intersects the North Hemisphere in latitude 78° 32′ N. and longitude 69° 08′ W. of Greenwich."

Silurian. By C. K. SWARTZ, W. F. PROUTY, E. O. ULRICH, and R. S. BASSLER. Maryland Geological Survey, Baltimore, Md., 1923. Pp. 794, pls. 67, figs. 27.

This recently published volume on the Silurian of Maryland maintains well the high standard set by the previous volumes of the Maryland Geological Survey.

Both the lithologic and faunal features of the Silurian rocks are shown with a wealth of detail in a series of sections by Dr. Swartz and Dr. Prouty. The more significant features of these sections are admirably summed up in a preliminary chapter by Dr. Swartz which includes two tables, one showing the formations, lithology, and faunal zones recognized, the other indicating the increase in thickness of the sections toward the east between Keyser, West Virginia, and the Delaware River, Pennsylvania.

The section on the correlation of the Maryland Silurian formations with those of other areas by Dr. Swartz includes two noteworthy features —a table showing the history of geologic nomenclature in Maryland and a tabular presentation of available knowledge concerning the range and distribution in eastern America of the Maryland Silurian fauna. The nomenclature used by eleven geologists, who during the past eighty-five years have recorded their studies of the Maryland Silurian rocks, is shown by the nomenclature table. The three stratigraphic subdivisions made by the Rogers brothers in 1836 have expanded to twenty-four divisions in the refined studies of Dr. Swartz and his associates. In the present state of geologic nomenclature such a table might well be considered a valuable, if not an essential, feature of any elaborate or comprehensive stratigraphic paper. Without such an aid the geologist who is not a specialist on the region or horizon discussed is likely, in the present stage of the development of stratigraphy, to find in many recent papers the highly localized nomenclature which refined work is now developing obscure. Dr. Swartz has, in this report, retained only two of the seven major divisions of the Silurian recognized by the Maryland Survey in 1897. This may lead the reader to speculate on the probable fate of some of the names used in the present report at the hands of future workers in the Maryland field.

The history of the various classifications of the Silurian in America is dealt with in a paper by Dr. Ulrich and Dr. Bassler. The references to the literature omit mention of some important papers relating to this subject. One of these by Cumings (*Handbook of Indiana Geology*) has

appeared too recently to have been included. This and certain other parts of the volume dealing with geological nomenclature and the time scale apply the views which Dr. Ulrich has advocated for a decade regarding the shifting of the Ordovician-Silurian boundary downward. Consequently the Silurian of much of this volume is not the Silurian of the textbooks nor of the reports of most other state, provincial, and government surveys of North America. It is the Silurian of most other authors and reports plus the part of the Ordovician known as the Richmond in the Ohio valley and the Queenston shale in New York-Ontario sections and the Juniata in Pennsylvania. This revision considerably more than doubles the thickness of the Silurian system in western New York.

The writer has somewhat the same interest in this alteration of the geological time scale which he would have in a proposal to lengthen the month of June by adding a couple of weeks from the month of July. If the State of Maryland wishes to have either a geological time scale or an almanac which differs radically from those generally used elsewhere that is her privilege, but the writer hopes that other states will not add to the confusion by following the example of some of the authors of this volume in redefining the Silurian system.

It must be added that it is not quite clear from the volume itself whether this revision of the Ordovician-Silurian boundary represents the official viewpoint of the Maryland Survey, or only a courtesy or privilege allowed two of the authors. A map labeled "The Silurian Formations of Maryland" which includes the Juniata formation and bears the State Geologist's name appears to support the first-named inference; but since it is incorporated in a report of Dr. Swartz, who adheres in his delimitation of the Ordovician-Silurian boundary to the customary usage, the reader closes the volume with a rather hazy idea as to just what the Maryland Survey's official attitude is on this question. The policy of leaving each author entire freedom in matters of nomenclature, which appears to have governed in the preparation of this volume, looks well from the standpoint of individual liberty but its weakness lies in the fact that it leads toward general chaos. The bewildering effect of two schemes of geological nomenclature in the same volume well illustrates the great need of geologists' getting together and using a nomenclature which is generally recognized.

Intimately connected with the revision of the Ordovician-Silurian boundary as defined by two of the authors of this volume, is the use of

the term Medinan which, as here used by Dr. Ulrich and Dr. Bassler, includes most of the Oswegan of Clarke and Schuchert plus the Richmond or uppermost Ordovician formation of most other authors. Those who wish to use the term Medinan without including formations in both the Silurian and Ordovician of most authors can use it as Dr. Swartz does in the present volume and as proposed by Cumings in the *Handbook of Indiana Geology*. The latter definition limits it to the beds between the Clinton (Hall, Genesee River section) and the Queenston, thus confining the name to the Silurian system as that term is used in Canada and most of the United States. At least five different meanings have been assigned to the term Medina or Medinan during the last ten years, not including such combinations as "Upper Medinan Albion formation" of the present volume. As nomenclature matters now stand, no other term unless it is Clinton can be more highly recommended to a geologist who, through lack of information or some other reason, finds it inconvenient to use a name with a precise meaning for a lower Silurian horizon.

Stratigraphic palaeontology presents no more baffling problem than the question of how to protect it from its friends. Until they are willing to use a common nomenclature the subject will become steadily a less exact science, if not a chaos, and fairly deserve the contempt of workers in the more exact sciences. If the history of geology from the days of Murchison and Sedgewick down to 1923 proves any one thing more clearly than others, it is that specialists may nearly always be counted on to completely disagree among themselves regarding the adjustment of nomenclature questions which concern other geologists quite as much as themselves. It seems therefore that the larger questions in geologic nomenclature such as the Ordovician-Silurian boundary should not be left indefinitely at the mercy of dialectic combat. In such cases the common sense of North American geologists should be enlisted in approving one of the two contentions or a possible third proposal. This could be secured by means of a committee representing all of the organizations or institutions concerned with the promotion of geological science in North America. While the decisions of such a committee would be binding on no one they would certainly furnish responsible survey heads, college professors, authors of textbooks and others who stand between the specialists and the public, a valuable index of general geologic opinion concerning any disputed point in geological nomenclature, and have in general a steadying influence on the fluctuations of geological nomenclature.

Half the volume is devoted to systematic palaeontology. The sections on the Ostracoda and Bryozoa are the work of Dr. Ulrich and Dr. Bassler. The other phylla have been treated by Dr. Swartz and Dr. Prouty.

The Cephalopod fauna is strikingly limited in the number of species and genera represented as compared with the richness of this fauna in such states as New York and Illinois. The Crinoidea, which furnish some unusual forms in Maryland, are not included in the scope of the work. The beautifully executed plates leave nothing to be desired in the way of illustrations.

The information given for each species regarding its horizon varies greatly in preciseness and consequently in value. Concerning the horizon of *Drepanellina confluens* n. sp., for example, no horizon or formation is given. In contrast with this case of negative information for which the authors are probably in no way responsible, "*Mastigobolbina micula*, Clinton—one hundred and two feet beneath the top of the Keefer sandstone" may be cited as an example of the precise and model statement of the horizon which concludes many of the specific descriptions. Contrasted with this model statement which gives for the fossil not only the formation but the exact location within it, and in addition to this definite information the opinion of the authors that these beds are the equivalent of some part of the New York Clinton, we find a great many cases where the horizon is cited as Clinton. This kind of a horizon citation gives merely the opinion held by the authors at a particular time concerning the correlation of the beds from which a given species is derived in terms of their definition of a time term during a particular year, or term of years, and fails altogether to supply information which would enable anyone to precisely locate it in the Maryland section. Since the attempt was made a few years ago to revise the meaning of the name Clinton by abandoning the meaning set for it in the later and best work of James Hall, it has lacked the definite meaning which it previously had in the New York Genesee section. The nomenclature used in this volume disregards the conclusions of Chadwick whose detailed stratigraphic work on the Clinton represents the latest work of the New York Survey on this terrane, and fails to accept Bassler's interpretation of the limits of Clinton as given in his *Bibliographic Index of Ordovician and Silurian Fossils*. The limits set for the Clinton by Ulrich and Stose in 1912 also appear to differ from those drawn for it in this volume. The mercurial character which the term has thus acquired in geological litera-

ture makes it therefore a very unsatisfactory horizon term when used alone.

The section on the morphology, classification, and occurrence of Palaeozoic ostracods, which is the work of the two foremost authorities in America on this group—Dr. Ulrich and Dr. Bassler—will be indispensable to all students of these fossils. It includes in addition to all that the title suggests an excellent discussion of methods of study. This extremely valuable contribution should go far toward encouraging students to take up the study of this somewhat neglected group of fossils.

The Maryland Survey deserves the thanks of all palaeontologists for the admirable style in which this much-needed volume has been published.

E. M. KINDLE

North American Later Tertiary and Quaternary Bryozoa. By FERDINAND CANU and RAY S. BASSLER. United States National Museum, Bulletin 125, Washington, 1923. Pp. i-vii+1-302, pls. I-XLVII, 37 figures in text.

This volume forms the concluding part of studies by the same authors on the Tertiary and Quaternary Bryozoa of North America. Like its predecessor[1] this one conforms to the high standard of scientific research established in this group of organisms. The Bryozoa are now in a position to be used by the geologist and stratigrapher as a valuable aid in the tracing and identification of marine sediments. If other groups of animals had been as thoroughly described and excellently pictured as this, many problems would be much easier of solution.

The major portion of the book consists of descriptions of Miocene and later fossil Bryozoa arranged according to the scheme of classification previously adopted. A catalogue of all papers dealing with the subject from 1841 (the first) to date is given with a synopsis of the contents. Also a bibliography of the literature of the Bryozoa, living and fossil, is given for the years 1899 to 1923. Literature prior to 1900 is contained in a previous publication.[2] The plates are made from a large number of individual photographs of striking beauty and clearness. The excellence of these pictures deserves warm commendation because they will be used probably more than any other part of the book, and because

[1] *North American Early Tertiary Bryozoa*, United States Nat. Mus. Bull. 106, 2 vols., 879 pp., 162 pls.

[2] Nickles and Bassler, United States Geological Survey, *Bulletin 173*.

there were technical difficulties of no small magnitude involved in their preparation.

Those who have engaged in similar detailed studies of other groups of animals can well appreciate the enormous amount of detailed work necessary to bring a publication such as this from the press. The necessity for checking and cross-checking of data, over and over again is depressing labor, yet in this case it seems to have been very thoroughly done. A thorough examination of the portions dealing with Pacific Coast species afforded very little opportunity for criticism. On page 14 the names *Porella cyclopea* and *Idmonea clarki* are listed from the Pleistocene of Santa Monica, California, but descriptions under these names do not appear in the body of the paper. The latter, however, is described under the name *Filisparsa clarki*.

G. DALLAS HANNA

Die Ichthyosaurier des Lias und ihre Zusammenhänge. By FRIEDRICH VON HUENE. Berlin. Gebrüder Borntraeger, 1922. Pp. 114, pls. 22.

A treatment of the Ichthyosauria as a whole has never been attempted before. The greater part of the literature on this interesting group of marine reptiles has been devoted to the description of specific forms (usually under the blanket name of "Ichthyosaurus"), and the taxonomy and phylogeny of the group have remained in a chaotic condition.

This deficiency is removed to a great extent by the present work of von Huene. Although dealing primarily with Liassic forms, those of other formations are reviewed and the taxonomy revised, while a large number of figures and a comprehensive bibliography add to the value of the volume. The stratigraphy of the Liassic ichthyosaur beds is discussed. Of especial interest is a description of the development of Stenopterygius from an 18-centimeter embryo to a 3-meter adult.

The division into longipinnate and latipinnate types, suggested by earlier writers, is developed. Von Huene believes that the entire order, except for a few early forms, may be divided into two phylogenetic series on this basis.

The author lists a large number of characters common to ichthyosaurs and Mesosaurus; but perhaps it might be better to await further fossil evidence before definitely uniting the two. Von Huene derives the common stem of ichthyosaurs and mesosaurs directly from the embolo-

merous amphibia. But the list of features in which Mesosaurus differs from the cotylosaurs, as cited by von Huene, seems insufficient evidence on which to revive the idea of a multiple origin of the reptiles.

A. S. ROMER

Die Familien der Reptilien. By FRANZ BARON NOPCSA. Fortschritte der Geologie und Palaeontologie. Heft 2. Berlin: Gebrüder Borntraeger, 1923. Pp. 210, pls. 6.

The rapid strides made in our knowledge of fossil reptiles during the past two decades have led to a number of attempts at a reclassification of the group. Nopcsa's arrangement contains many interesting features but, like its predecessors, must be regarded as provisional in many respects.

Some twenty-five "types" are selected, around which are centered the discussions of the groups to which they belong. More general questions of relationships are treated in a later section. Twenty-one ordinal groups are established, arranged in ten superorders. A discussion of reptilian footprints is a novel feature.

From the nature of the subject, many of the conclusions are, of course, highly debatable. The group "Rhizosauria," root-reptiles, is established for the reception of Datheosurus, Eosauravus, and Sauravus. But the former is incompletely known, and the two latter are probably lepospondylous amphibians. The comparison of expanded cotylosaur ribs or rib plates with Eunotosaurus and chelonian ribs and plates is untenable, as a consideration of the muscular relations of the forms shows.

The use of "Pelycosauria" for all the American Permo-carboniferous forms with a lateral temporal opening is conservative and perhaps justified. But the superfamilial separation of Ophiacodon from Theropleura and Diopaeus, which are practically indistinguishable from it except in the temporal region, seems unnatural. The inclusion of the Thalattosauria in the pelycosaurs seems somewhat rash.

The treatment of the mammal-like reptiles is radically different from that now generally accepted. Cope's term Theromorpha is used exclusively for South African forms, in sharp contrast to the usage of Williston and other writers. The many resemblances noted by Broom, Watson, and even many earlier authors between pelycosaurs and the South African forms are briefly dismissed; the homology of the temporal openings in the two groups is denied. The dicynodonts and dromasaurs are grouped together as "Chainosauria" on not very obvious grounds.

The two arched reptiles are placed in three superorders. In the first (Diaptosauria) are placed the Rhynchocephalia and Thecodontia. (Proterosaurus finds a resting place in the latter.) This association seems unfortunately conservative. Sphenodon and its allies at present appear to be only remotely related to the other diapsids, while the thecodonts are unquestionably ancestral to the dinosaurs, which comprise Nopcsa's second diapsid superorder, and to the pterosaurs and crocodiles. The last two are united in the superorder Praepubci, for no reason other than the supposed presence of a prepubis in both groups. But it is quite doubtful whether the Crocodilia have a prepubis; the muscular argument cited by Nopcsa can be disproved. Otherwise there seem to be no features common to the crocodilia and flying reptiles other than primitive archosaurian characters.

The work shows evidences of a very careful consideration of the recent literature, and, no matter what opinions may be held as to some of the conclusions, furnishes stimulating reading for anyone interested in reptilian phylogeny.

A. S. ROMER

Sammlung geologischer Führer. Berlin: Gebrüder Borntraeger, 1923. Vol. XXII, *Die Westtiroler Zentralalpen*, by WILHELM HAMMER. Pp. 150, figs. 22, plate 3. Price $0.75. Vol. XXIII, *Helgoland und die umliegenden Meeresgründe*, by OTTO PRATJE. Pp. 115, figs. 8, maps 4, profiles 2.

Two new geological guidebooks in the extensive series being published by Gebrüder Borntraeger.

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